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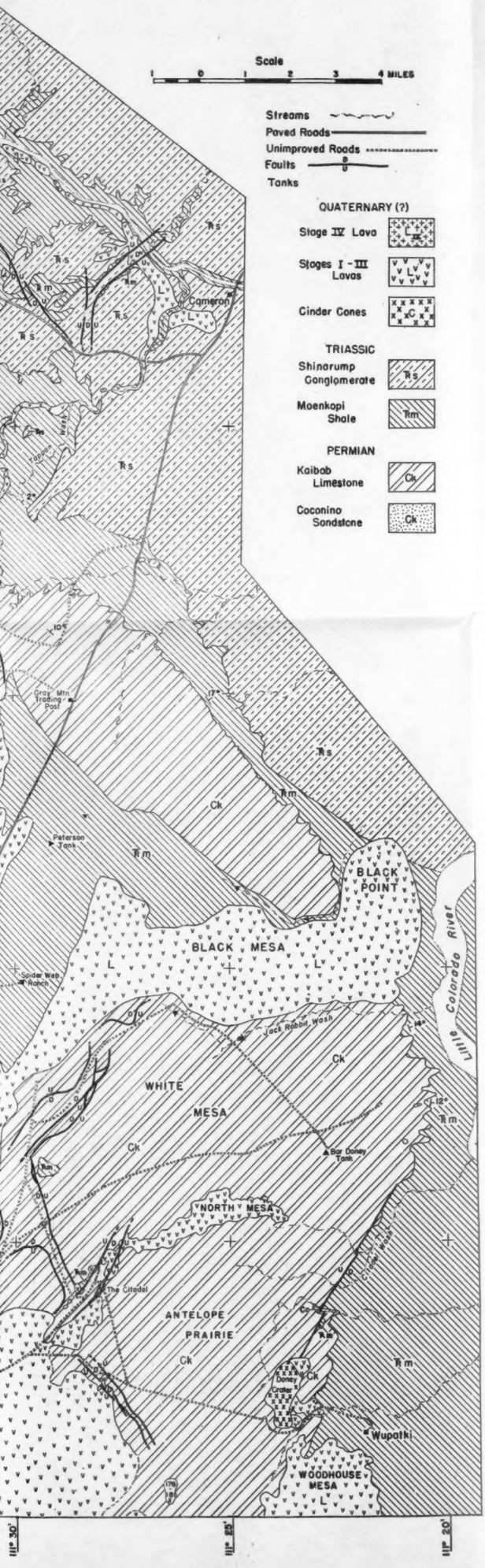
GEOLOGIC MAP OF THE CAMERON-WUPATKI REGION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Moenkopi Shale
PERMIAN
Kaibab Limestone
Coconino Sandstone
Cameron
Gray Mtn Trading Post
Peterson Tank
BLACK POINT
BLACK MESA
Spider Web Ranch
Jack Rabbit Wash
Little Colorado River
WHITE MESA
Bar Doney Tank
NORTH MESA
The Citadel
ANTELOPE PRAIRIE
Doney Crater
Wupatki
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111° 30'
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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GEOMORPHOLOGY AND STRUCTURE OF THE EAST KAIBAB MONOCLINE, ARIZONA AND UTAH

BY DONALD L. BABENROTH* AND ARTHUR N. STRAHLER

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ABSTRACT

The East Kaibab monocline, a compound flexure downthrown 2000 to 5000 feet on the east side, extends from the San Francisco Peaks volcanic field, Arizona, north 150 miles to Bryce Canyon, Utah. South of the Grand Canyon the monocline has several branches which make abrupt changes of strike. The main flexure west of Cameron trends east, while other segments strike northwest,

north, and northeast. With many small grabens, single faults, and rows of cinder cones the flexures form a structural pattern of three intersecting trends.

The main monocline, paralleled by the Butte fault, runs north across the eastern end of the Grand Canyon. A minor branch, paralleled by the Cremation fault, runs northwest into the Grand Canyon from Grandview Point. The Colorado River follows the strike of these major structures and may have had a subsequent origin in a belt of weak Triassic strata swinging south around the plunging Kaibab arch. The Little Colorado may have developed later in a similar belt at the foot of the East Kaibab monocline.

North of the Grand Canyon the monocline forms the eastern Kaibab Plateau margin and for part of this distance is a double or triple flexure. In Utah the monocline is a single great fold in Jurassic and Cretaceous strata. At Canaan Peak the Eocene Wasatch formation unconformably overlies beveled Upper Cretaceous strata in the monocline, indicating Laramide flexing.

INTRODUCTION AND ACKNOWLEDGMENTS

The East Kaibab monocline, one of several major north-trending structures of the western Colorado plateaus, extends from the San Francisco Peaks volcanic field, Arizona, north across the Grand Canyon and along the east side of the Kaibab Plateau to Bryce Canyon National Park, Utah. The monocline has never been described in detail as a unit, although it has been intensively studied within the Grand Canyon and in southern Utah. The present paper gives a detailed geologic description of the East Kaibab monocline and immediately adjacent areas, discusses the geomorphology of the same areas, and offers a hypothesis of origin for the Colorado River in the eastern part of the Grand Canyon.

The project was begun by the senior author as the subject for a doctoral dissertation at Columbia University. Most of the field data were gathered by him during the summers of 1939 and 1940. After the death of the senior author, in 1940, the junior author, who had served as field assistant, agreed to complete the project and received from the Penrose Bequest of The Geological Society of America a grant making possible a third summer of field work in 1941. All writing, compilation, and drafting of figures and maps has been done by the junior author who has also contributed original material to the discussion of geomorphic problems.

The writers wish to acknowledge the generous assistance of Mr. Edwin D. McKee in suggesting the problem, following closely the progress of the work, and making available facilities of the Grand Canyon National Park and the Museum of Northern Arizona. He has also read the manuscript and offered many helpful suggestions. The late Professor Douglas Johnson directed the work and visited the area. Professor Walter H. Bucher critically read the manuscript. Valuable field assistance was given by H. Holt Riddleburger and E. D. Koons. Members of the Columbia Seminar in Geomorphology of the years 1938–1941 offered many useful criticisms.

GENERAL DESCRIPTION

The East Kaibab monocline is the westernmost of the north-trending monoclines which form the Plateau province into steps (Fig. 1). Because all these monoclines are downthrown on the east, each plateau block becomes successively higher from east to west. The East Kaibab monocline thus raises the highest of the blocks, the Kaibab Plateau, to elevations over 9000 feet. West from the Kaibab Plateau is a succession of great north-trending faults with downthrow on the west. Thus the plateau blocks become progressively lower as one goes west from the Kaibab Plateau, the keystone block of the region.

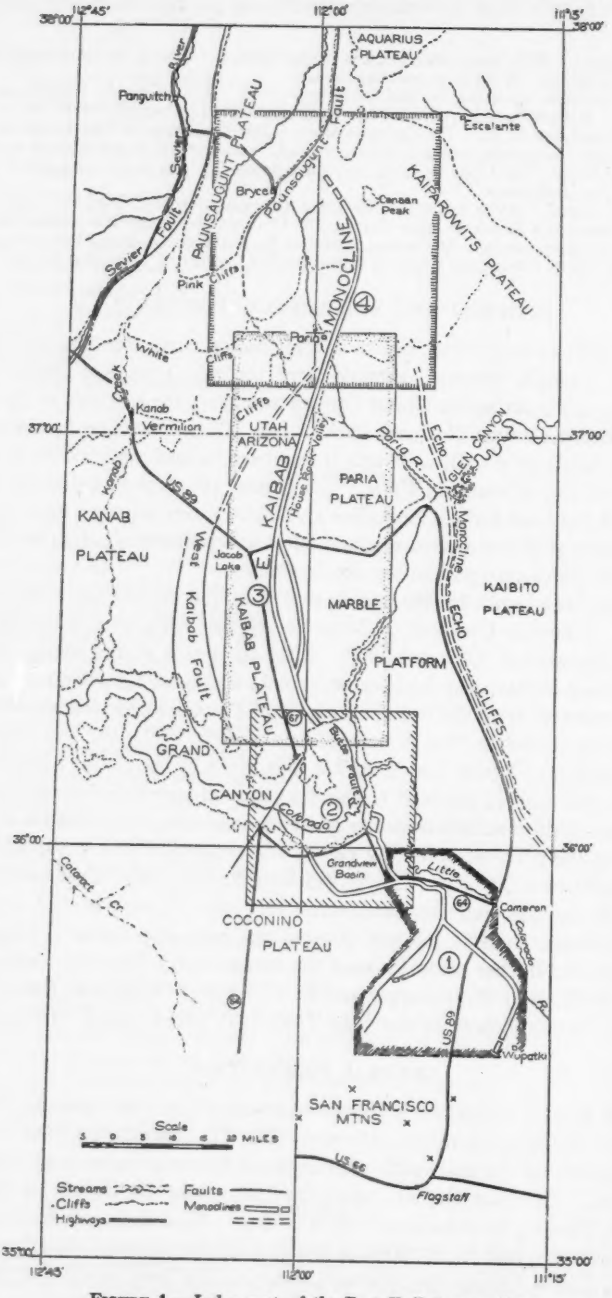


FIGURE 1.—*Index map of the East Kaibab monocline*

Portions outlined represent areas included in geologic maps of the four divisions of the East Kaibab monocline: (1) Cameron-Wupatki division, (2) Grand Canyon division, (3) Kaibab Plateau division, and (4) Paria division.

The monocline is about 150 miles long with a displacement of 2000 to 5000 feet. Its southern terminus is near the northern fringe of the San Francisco Peaks volcanic field, Arizona. North to the Grand Canyon the monocline forms the eastern margin of the Coconino Plateau, along which the strata are flexed down to form the southern end of the Marble Platform.

South of the Grand Canyon the monocline characteristically branches into flexures separated by platforms of flat-lying strata. The flexures locally show great divergences from the general north trend; the most noteworthy example is the main part of the monocline west of Cameron which strikes almost due east and has a throw of 2000 feet. Many single faults and grabens lie on or near the southern branches.

The monocline continues north across the eastern end of the Grand Canyon, which has been opened up along the strike of the main fold. North of the Canyon the monocline steps down the strata 3000 feet from the Kaibab Plateau to the Marble Platform; it then continues north into Utah, with throw increasing to over 5000 feet, and dies out east of the Paunsaugunt Plateau near Bryce Canyon.

The monocline is here treated in four major "divisions." The separation, largely geographic, is employed solely to facilitate description, and the structure does not necessarily show any important change where the monocline passes from one division to the next. From south to north there is: (1) the Cameron-Wupatki division, (2) the Grand Canyon division, (3) the Kaibab Plateau division, and (4) the Paria division. Each includes smaller "sections," most of which represent branches of the monocline or segments separated by abrupt changes in strike, and are therefore structural units.

PREVIOUS WORK

Although many writers have mentioned the monocline briefiy, only a few have offered important contributions to an understanding of the areal and structural geology of the fold or its geomorphic development. Newberry (1861, p. 61), on Ives' exploring expedition of the Colorado River, first described but did not name a downflexing of Carboniferous strata between the Coconino Plateau and the Little Colorado River valley. Powell (1875, p. 186, 192–193) described the structure along the east side of the Kaibab Plateau and named it the East Kaibab fault or East Kaibab fold. Gilbert (1875, p. 52–53) added details concerning the same area. Dutton (1880, p. 32–33) first noted the general extent and character of the monocline at its northern end and later (1882, p. 177–178, 183–186) gave an excellent though still generalized description of the East Kaibab monocline north of the Grand Canyon. He was unable to examine the structure south of the Canyon but described what he could see of it from Cape Final on the north rim. He considered the folding to be early Pliocene, following establishment of the Colorado River.

Walcott (1890; 1895) made the only previous intensive areal and structural study of the monocline in Arizona. His work was limited to the eastern end of the Grand Canyon because he was principally concerned with the Butte fault lying along the axis of the monocline in the ancient rocks of the canyon bottom. He was the first to suggest that the monocline is relatively ancient, formed before extensive erosion of the region.

Davis (1901, p. 164–166) pointed out further details of the monocline south of the Grand Canyon and first suggested that the development of the Colorado and Little Colorado rivers was controlled by these structures. Johnson (1909, p. 140–141) noted numerous minor faults near the southern terminus of the monocline. Robinson (1913, p. 34–37) described the monocline in the vicinity of Coconino Point and the south rim of the Grand Canyon, but his most important contribution was a description of the flexure at Black Point southeast of Cameron. Gregory (1914) prepared a rough geologic map showing some of the larger structures of the region around Cameron. Gregory and Moore (1931), by describing and mapping the northern end of the monocline in the Kaiparowits region, greatly advanced our knowledge of the flexure in that area. Darton (1910; 1923; 1924; 1925) has compiled generalized maps showing areal geology, structure, and topography of the monocline in Arizona.

CAMERON-WUPATKI DIVISION

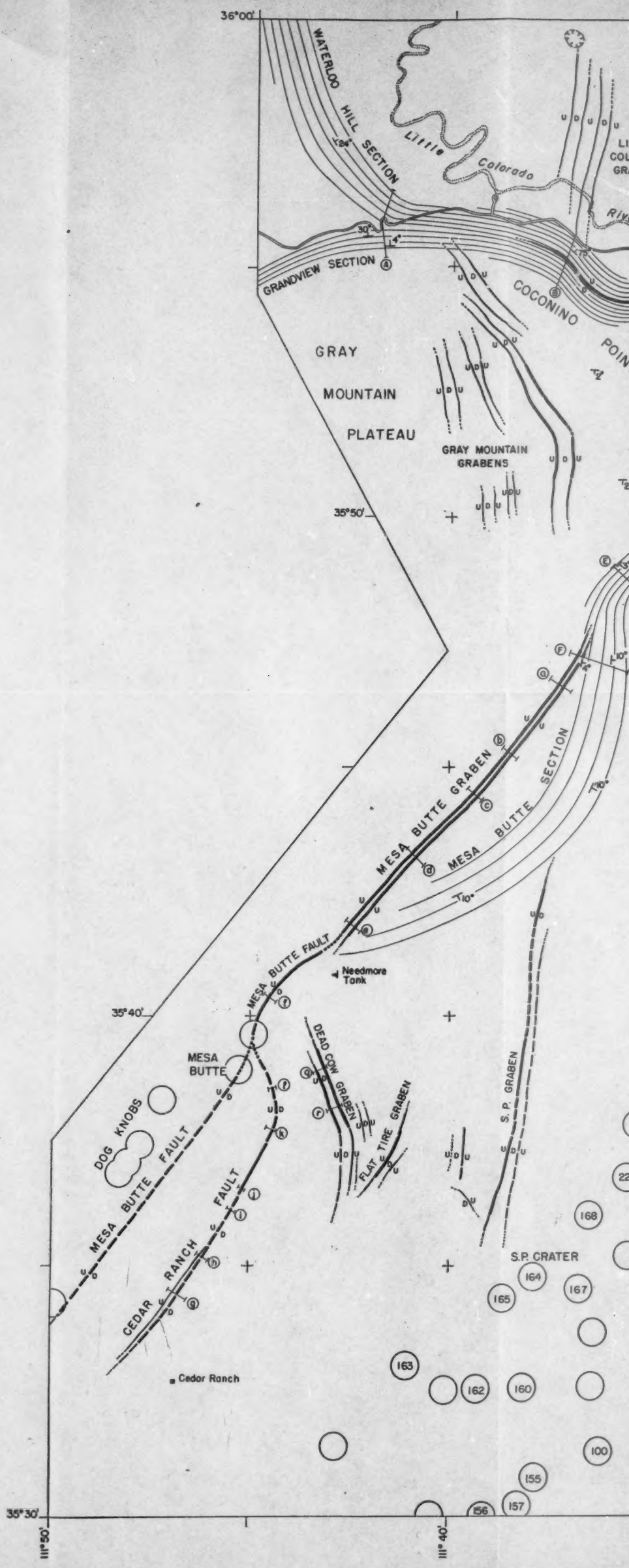
GENERAL STRUCTURE

This division lies south and west of the Little Colorado River between Cameron and the Wupatki National Monument (Fig. 1; Pls. 1, 2). Elevations range from slightly over 7000 feet on the west to less than 3500 feet in the Little Colorado valley. Most of the area is sparsely forested with juniper and pinyon, but large portions at the lower elevations are devoid of trees. U. S. Highway 89 leading from Flagstaff to southern Utah and State Highway 64 leading to the south rim of the Grand Canyon from Cameron provide easy access to some of the important geological features.

The monocline in the northwest Cameron-Wupatki region is a single sharp locally east-trending flexure. This part is called the Coconino Point section after Coconino Point, a sharp promontory formed as the monocline turns abruptly south. Near the center of the area the monocline bifurcates. The Black Point section trends southeast to Black Point on the Little Colorado River, then turns sharply southwest and passes under lavas of the Doney Craters. The other branch continues southwest and dies out near Mesa Butte. All the folds are downthrown on the north or east sides which causes the strata to step down from higher elevations on the west and south toward the structural depression of the Little Colorado valley.

Permian and Triassic formations surface most of the Cameron-Wupatki region, although Quaternary lavas and cinder cones of the San Francisco Peaks volcanic field (Colton, 1937) cover much of the southern part. The most important and widespread sedimentary rock is the magnesian limestone of Permian Kaibab formation. This highly resistant rock, from which the weak overlying Triassic beds are easily eroded, gives rise to large areas of stripped structural surface. The East Kaibab monocline is surfaced almost entirely by the Kaibab limestone throughout the entire length of the fold in Arizona. The weak Triassic formations are now extensively preserved only near the Little Colorado River, structurally the lowest part of the region and therefore most protected from erosion.

The geologic map (Pl. 1) is based on controlled mosaics of the Soil Conservation Service on the scale of 1 inch to a mile. Geologic contacts and structures were examined in the field and traced from the mosaics. No attempt was made to



STRUCTURE MAP OF THE C

STRUCTURE MAP OF THE CAMERON-WUPATKI REGION NORTHERN ARIZONA

Scale
1 0 1 2 3 4 MILES

FAULTS
Throw 0 to 25 ft.
25 to 100 ft.
Over 100 ft.
Fault buried under lava
Probable fault

EAST KAIBAB MONOCLINE

Each space between lines represents 200 feet of throw. Lines are neither contours nor form lines.

Structure sections

LITTLE COLORADO GRABENS
River
COCONINO POINT
POINT SECTION
Grand Canyon
CAMERON GRABEN
Cameron
Highway
POVERTY TANK GRABEN
Poverty Tank
Gray Mtn. Trading Post
CAMPBELL & FRANCIS GRABEN
Campbell Francis Tank
BLACK POINT SECTION
BLACK POINT
Little Colorado River
BLACK MESA
WHITE MESA
WHITE MESA GRABEN
E GRABEN
CITADEL
The Citadel
Highway U.S. 89
DONEY FAULT
DONEY CRATER
Wupatki
HULL CANYON GRABEN
226
169
225
168
167
100
178
181
111°30'
111°20'

P OF THE CAMERON-WUPATKI REGION

represent topography as very few elevations were available. The structure map (Pl. 2) shows only flexures, faults, and cinder cones. Colton (1937) mapped and classified lava flows and cones of the region.

The Toroweap formation (McKee, 1938), although a recognized stratigraphic unit, was included in the Kaibab formation which it underlies. Most of the distinctive features which make it easily recognizable in the type locality far to the west are here absent, and the upper contact is not easy to locate.

So far as the writers are aware, none of the monocline segments and branches nor the single faults and grabens have been named previously. Most names used in this paper are from near-by geographic features.

MONOCLINES AND LARGER F ULTS

Coconino Point section.—The Coconino Point section is the most prominent structural and topographic feature within the Cameron-Wupatki area (Pls. 1, 2). Its western end begins near the junction of Gray Mountain Truck Trail and Grand Canyon Highway. There two branches of the monocline, the Grandview and the Waterloo Hill sections, join to form a single large fold (Pl. 3, fig. 2) which continues eastward toward Cameron for 8 miles, then turns sharply south. This abrupt bend produces Coconino Point which overlooks the Little Colorado valley (Pl. 3, fig. 1). The fold continues south for 3 miles, then makes another sharp turn southwest. Four miles farther the section ends where the fold changes character and strike.

Structure sections of the Coconino Point monocline were made at about 3-mile intervals (Fig. 2 A–E). This part of the monocline is characterized by relatively great displacement within small horizontal distance. The maximum displacement measured was 2300 feet in a horizontal distance of a mile (Fig. 2 B). The throw is greatest within 4 miles in either direction from Coconino Point and diminishes to 1200 and 1500 feet at the west and south ends respectively.

Maximum observed dip of the strata in this part of the monocline is 89° (Fig. 2 B, C) within a few miles of Coconino Point where the throw is likewise greatest. Where the highest dip prevails the flexing is extremely sharp in the Moenkopi shale. In some road cuts the Moenkopi bedding is almost vertical on one side of the road, but nearly horizontal on the opposite side, yet there is no evidence of faulting. The incompetence of this formation may well explain the sharp bending, for similar conditions were found nowhere in the massive limestones of the Kaibab formation.

The Gray Mountain Plateau, southwest of the Coconino Point section, is part of a much larger unit, the Coconino Plateau, which lies south of the Grand Canyon. The Gray Mountain Plateau is highest, about 7000 feet, at its northern edge just at the crest of the monocline. It is capped by the Kaibab formation which dips about 2° S. This may be responsible in part for the diminishing displacement of the monocline south from Coconino Point.

Northward from the base of the Coconino Point fold is the Marble Platform, elevation about 5000 feet, into which the Little Colorado River has cut a deep, narrow canyon. Here again the stripped top of the Kaibab limestone forms the surface. Near the base of the monocline a superficial strip of Moenkopi shale is still preserved, but the Kaibab formation reappears near the brink of the canyon.

The monocline has been deeply breached, and nowhere along its axis do the

uppermost Kaibab limestone beds extend unbroken from the base to the crest. From a great distance the fold looks like a plateau escarpment, doubtless the reason why Dutton (1882, p. 185), viewing it from Cape Final on the north rim of the

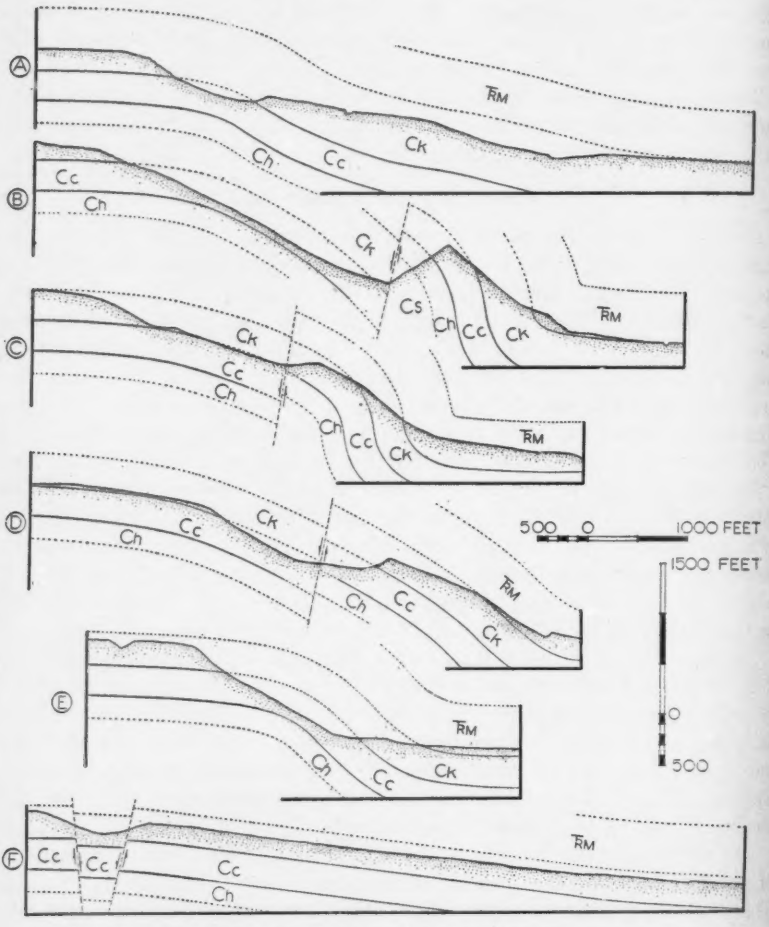


FIGURE 2.—*Structure sections of Coconino Point and Mesa Butte sections*

Marginal letters refer to position of sections as indicated on Plate 2. Cs—Supai fm. Ch—Hermit fm. Cc—Coconino fm. Ck—Kaibab fm. Trm—Moenkopi fm.

Grand Canyon, described it as a fault. Closer view reveals clearly a great series of flatirons along the base of the cliff (Pl. 3). Erosion has exposed 200 or 300 feet of Coconino sandstone along most of the fold axis. Locally (Fig. 2 B) the Hermit and Supai formations have probably been exposed although they are now concealed by talus. These are not mapped.

On the structure sections (Fig. 2 B–E) a fault is indicated along the axis of the

FIGURE 1.—COCONINO POINT SECTION OF EAST KAIBAB MONOCLINE
Viewed from the Marble Platform. Coconino Point at extreme left.

FIGURE 2.—PANORAMIC CONTINUATION OF FIGURE 1
Showing Waterloo Hill section of monocline nearest observer in middle and right, and Grandview section of monocline on distant skyline at right.

EAST KAIBAB MONOCLINE AND MARBLE PLATFORM

FIGURE 1.—MESA BUTTE GRABEN
View northeastward down the length of the graben from a point near its southwestern end.

FIGURE 2.—VIEW NORTHWARD ALONG THE DONEY FAULT
From the summit of Doney Crater, showing resequent fault-line scarp of Kaibab limestone.

FIGURE 3.—EAST WALL OF WHITE MESA GRABEN
Graben floor in upper left of view. Graben wall, in foreground, is a sharp monoclinal flexure with buckle-ridge at the crest. Rock is the Kaibab limestone.

monocline but having opposite throw. Its displacement was calculated to be 200 to 300 feet, with the south or west sides downthrown. It was not observed in the field because of its location, partly in the Coconino sandstone where cross-bedding confuses the secondary structures and prevents recognition of horizons, and partly along longitudinal axial valleys breaching the monocline where talus and alluvium obscure much of the bedrock. The fault is inferred to be post-monoclinal and may have been produced after much of the overlying rock was removed, thus accounting for the strata breaking rather than bending as when buried under a great thickness of rock.

Mesa Butte section.—Six miles south of Coconino Point the monocline changes abruptly from a sharp flexure of great throw to a broad, gentle fold of much less displacement. This point marks the north end of the Mesa Butte section. Southward the monocline curves gradually. At the southern end where the fold dies out, it strikes southwest. This part of the monocline resembles a crescent with the convex or downthrown side facing southeast.

Displacement in the Mesa Butte section measured 750 feet near its northern end (Fig. 2 F). The monocline is about 1½ miles wide. Maximum dip recorded is 10°. The flexure disappears 3 miles northeast of Mesa Butte.

In the Mesa Butte section the Kaibab limestone forms the surface of the fold and of the plateaus on both sides of it. The fold has not been breached. Viewed from the base it resembles a broad ramp of limestone with convex upper surface scored by numerous small canyons. Along the base lava flows have inundated most of the lower plateau surface and rest against the rising limestone layers.

Black Point section.—The Black Point section stems from the Coconino Point section about 5 miles south of Coconino Point (Pls. 1, 2). The junction of the two folds is extremely abrupt; the Black Point fold abuts the Coconino Point fold at right angles. Despite this, fusion of the two monoclines is accomplished by the strata bending without faulting.

The Black Point monocline extends southeast 14 miles to Black Point, a mesa of lava which flowed across the truncated fold. Near Black Point the monocline turns toward the south. Five miles south of Black Point it makes a sharp turn to somewhat west of south. Here the displacement diminishes rapidly, and a fault takes up most of the throw.

This section is characterized by relatively small displacement and low dips. Transverse structure sections were made at intervals of from 2 to 4 miles along the fold (Fig. 3). At its western end the flexure is broadest—about 3 miles—and the dip lowest, nowhere exceeding about 8°. Farther east toward Black Point the fold narrows to about 1½ miles, while the maximum dips increase to 20° (Fig. 3 B, C). Here the displacement exceeds 1000 feet, the greatest anywhere along this section. Four miles south of Black Point the throw is only 500 feet. The Doney fault, which lies along the monoclinal axis and has the same side downthrown, first appears about 6 miles south of Black Point. The monocline nevertheless continues, although combined with the fault, to a point under the Doney Craters (Pl. 4, fig. 2). South of these the fault has disappeared, but the flexure is still preceptible in the beds and continues southward under lavas.

The Black Point section is represented by the smooth, unbreached upper surface

of the Kaibab limestone. The weak Moenkopi shale has been cleanly stripped away to the fold base where low hogbacks and cuestas remain. Explanation of the youthful appearance of the monocline here as contrasted with the Coconino Point section is that the former lies at a much lower elevation and has only recently been

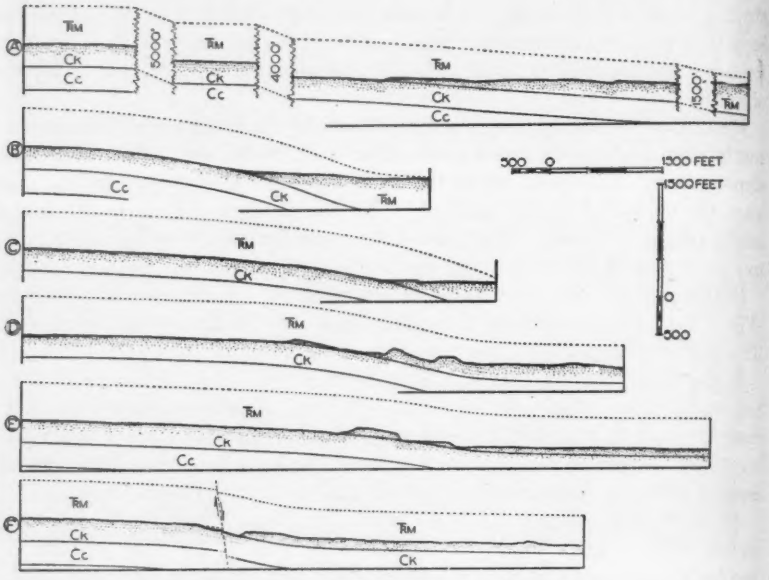


FIGURE 3.—*Structure sections of Black Point section*

Marginal letters refer to position of sections as indicated on Plate 2. Cc—Coconino fm. Ck—Kaibab fm. Trm—Moenkopi fm.

stripped of Mesozoic beds. At Black Point the lava flow which lies across the monoclinal axis protects it from erosion. This proves the monoclinal flexing to be of pre-lava (pre-Quaternary) date, a calculation which, however, is not of great significance because the extensive erosion indicates a much earlier date of folding.

Southwest of the Black Point flexure is a large area of nearly horizontal strata. Parts of this area are floored by the stripped Kaibab limestone, others by thin residual layers of Moenkopi shale, and the remainder covered by lava flows, some of which are eroded at their margins to produce mesas. Northeast of the Black Point flexure the Triassic Moenkopi, Shinarump, and Chinle formations, in that order, are preserved in the Little Colorado valley.

Mesa Butte graben.—One to 2 miles northwest of the Mesa Butte section a graben trends roughly parallel with the monocline. It is directly in line with the southern part of the Coconino Point section of the monocline with which it connects. The graben is about a quarter of a mile wide and 10 miles long. It terminates near the place where the Mesa Butte flexure dies out. The graben may be pictured in plan view as a trough connecting the ends of the crescent made by the monocline. From

its southern end the westernmost of the two bounding faults continues as a single fault, described elsewhere as the Mesa Butte fault.

Structure sections across the Mesa Butte graben were made at 1- to 2-mile intervals (Fig. 4). The two faults are about 1000 feet apart. The throw, where it

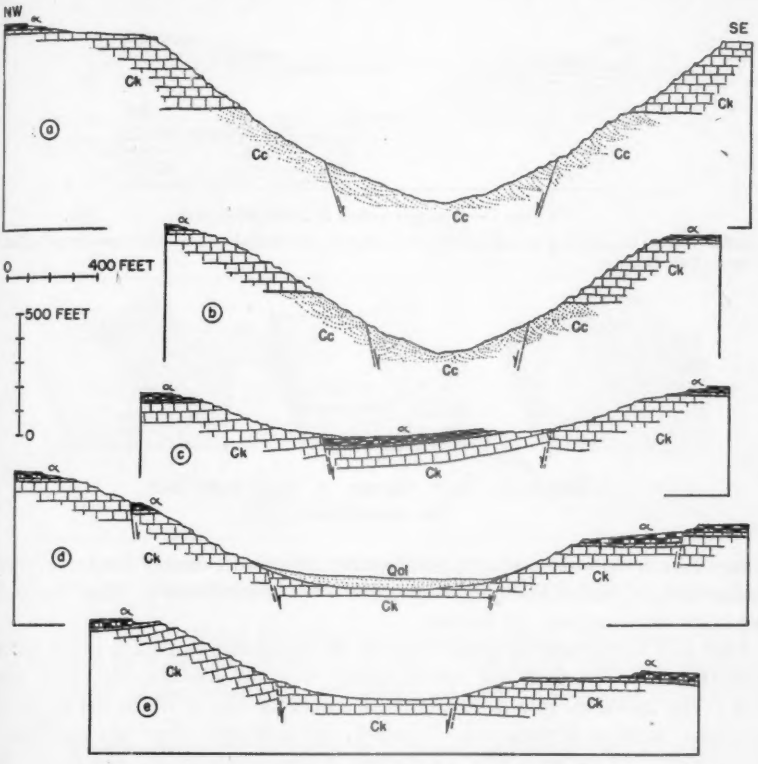


FIGURE 4.—*Structure sections of Mesa Butte graben*

Circled letters refer to position of sections as indicated on Plate 2. Cc—Coconino fm. Ck—Kaibab fm. α—Alpha member of Kaibab fm. Qal—Alluvium.

could be measured, is about 250 feet but may be considerably more elsewhere. The graben is now occupied by stream valleys wider and deeper than the original structural trough (Pl. 4, fig. 1). At its northern end is a canyon nearly 500 feet deep which becomes shallower to the south. Along the more deeply eroded northern part the Coconino sandstone is exposed.

Evidence is lacking for dating the graben faulting. That the faults are not very recent is attested to by dissection of the valley walls and deep erosion into the floor. Possibly the faulting is of relatively ancient date, and rapid removal of the weak overlying Triassic beds resulted in reappearance of a fault trench on the resistant surface of the Kaibab limestone, which in turn was slowly dissected and deepened.

If so the limestone walls would be resequent fault-line scarps rather than simple fault scarps.

Mesa Butte fault.—The Mesa Butte fault, described briefly by Johnson (109, p. 140–141), is a southwest continuation of the western wall of the Mesa Butte graben and is thus downthrown on the southeast. Three miles southwest of the

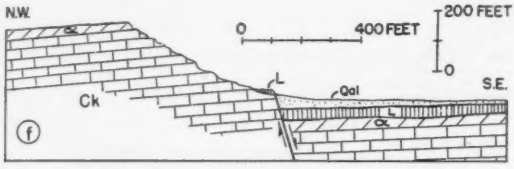


FIGURE 5.—*Structure section of Mesa Butte fault*

Letter (f) refers to position of section as indicated on Plate 2. Ck—Kaibab fm. α—Alpha member of Kaibab fm. L—Lava. Qal—Alluvium.

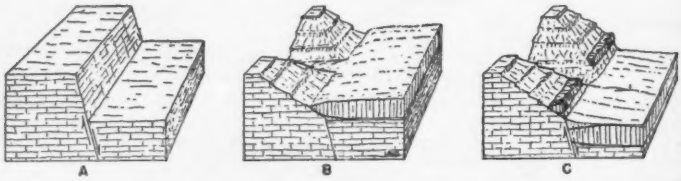


FIGURE 6.—*Block diagrams of Mesa Butte fault*

Show supposed history.

graben mouth the fault passes under Mesa Butte, a cinder cone of the San Francisco Peaks volcanic field. The fault was traced 7 miles southwest of Mesa Butte, but the exact terminus was not located.

That part of the fault northeast of Mesa Butte appears as a scarp about 250 feet high (Fig. 5). The fault line curves slightly with the concave side facing southeast. The upthrown side is of Kaibab limestone on top of which the uppermost, or Alpha, member is preserved. Strata in the upthrown block dip northwest at a low angle. At the scarp base a large area of alluvium obscures bedrock, but from the presence of lava farther out from the base, and elsewhere along the fault line, bedrock at the cliff base is thought to be lava.

Three small patches of lava cling to the limestone scarp about 20 feet above its base. They seem to represent remnants of a flow which once stood against the scarp at that level. Since the scarp is still highly rectilinear, modified only by narrow V canyons incised into it, and a study of the region shows no comparable erosion, it is unlikely that a great sheet of lava has been eroded from the level of these small patches down to the present surface at the scarp base. The explanation illustrated (Fig. 6) seems better suited to the facts. A shows the first scarp of Kaibab limestone. Whether it was a fault scarp, or a fault-line scarp freshly exhumed from under weak Triassic beds, cannot be determined. B shows the scarp reduced to an angle of about 30° by weathering along the entire face, while V canyons were cut into the scarp. Then a lava flow inundated the downthrown block and flowed up

against the scarp and into the canyon mouths. C shows renewed faulting along the same fault line causing a small strip of lava to be raised above the base of the scarp. Today only a few patches of lava remain, the rest having been eroded. No lava was found in the canyon mouths, but this is expectable because stream action would

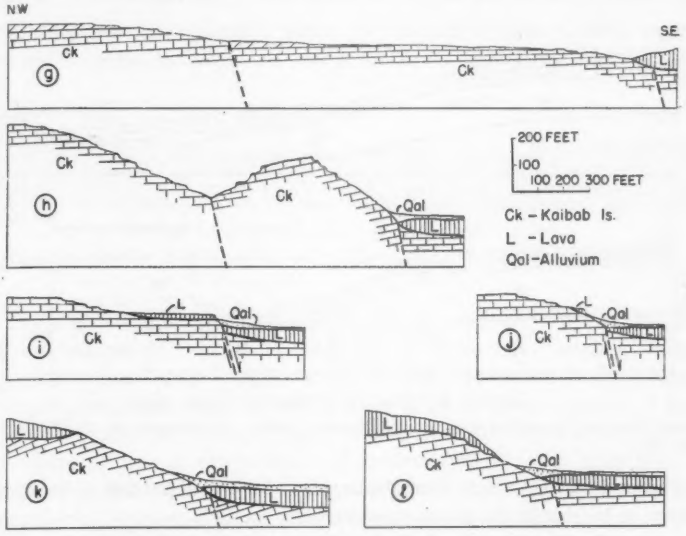


FIGURE 7.—*Structure sections of Cedar Ranch fault*
Circled letters refer to position of sections as indicated on Plate 2.

quickly cut it away. If the foregoing explanation is correct, the present scarp is a combination of fault or fault-line scarp with a fault scarp. Date of the earlier faulting cannot be determined, but the second faulting is Quaternary since it cuts the lava.

Southwest from Mesa Butte the fault is represented by a rectilinear scarp of Kaibab limestone against which rest lava flows. The fault line is therefore buried and is not mapped, but its presence is clearly indicated by the topography. Narrow V notches cut in the limestone scarp by streams have been invaded by the lava. No suggestion of recent faulting was seen.

The Dog Knobs, a series of small cinder cones, lie along a line a mile northwest of the Mesa Butte fault and parallel to it. Orientation of the cones suggests a line of weakness parallel to the fault.

Cedar Ranch fault.—The north end of the Cedar Ranch fault is under a small cinder cone just north of Mesa Butte and is probably a branch of the Mesa Butte fault. It runs southeast for about a mile from Mesa Butte, then curves south and southwest approximately parallel to the southern end of the Mesa Butte fault. Like the latter, the Cedar Ranch fault is downthrown on the southeast and is represented by a southeast-facing Kaibab limestone scarp against which lavas have flowed. It is, however, somewhat more complex in form and history (Fig. 7).

At its southern end, about 2 miles north of Cedar Ranch, the total displacement is about 200 feet distributed through two faults about a quarter of a mile apart. Lava flows cover the scarp base. No evidence of recent movement along this part of the fault was seen. Four miles north from Cedar Ranch the upthrown or northwestern block, as well as the downthrown one, was inundated by lavas so that

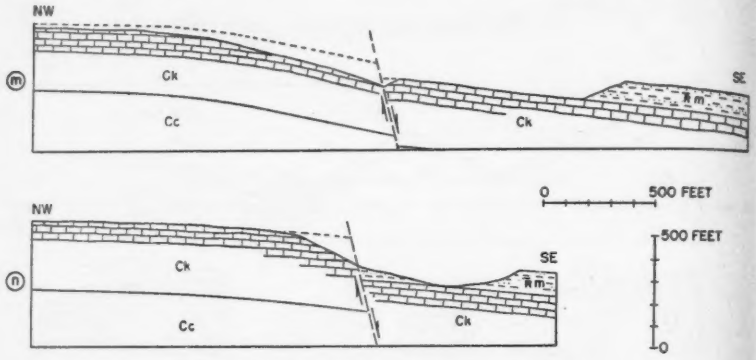


FIGURE 8.—*Structure sections of Doney fault*

Circled letters refer to position of sections as indicated on Plate 2. Cc—Coconino fm. Ck—Kaibab fm. Tm—Moenkopi fm.

from this point northward to Mesa Butte only a few small patches of limestone along the edge of the upthrown block show through the lava.

For about 2 miles along the fault line, beginning at a point 4 miles north of Cedar Ranch, there is evidence of recent movement along the fault line. Here a strip of lava, not more than 10 feet thick, clings to the limestone scarp about 20 feet above its base (Fig. 7, I, J). The lava strip is clearly not an erosional remnant since it is continuous along the scarp. The occurrence is similar to that described by Johnson (1909, p. 155–156, Pl. 21) for the Hurricane fault, but on a much smaller scale. The series of events is also similar to that outlined for the Mesa Butte fault north of Mesa Butte. The first faulting was of pre-lava date, but whether the upper part of the present scarp is a fault scarp or fault-line scarp cannot be determined. Then lava flowed up against the scarp base and was followed by a small, relatively recent movement along part of the old fault line, locally increasing the height of the scarp.

The Cedar Ranch scarp is almost entirely buried under lavas near the northern end, where it curves northwest into Mesa Butte, but is still a prominent topographic feature (Fig. 7 K, L). No evidence of recent faulting was found at this end.

Doney fault.—The Doney fault, along the axis of the Black Point section, first appears about 5 miles north of the Doney Craters. It runs under the cones but was not observed to emerge south of them. It is downthrown on the east, as is the monocline upon which it is superposed (Fig. 8). Maximum displacement observed on the fault is 200 feet.

The fault is seen as a scarp of Kaibab limestone in which the beds show clearly the monoclinal bending (Pl. 4, fig. 2). The Moenkopi formation covers much of

the downthrown block up to the fault line but in places is stripped back for half a mile. A mile north of Doney Craters a particularly deep canyon in the upthrown block has cut into the Coconino formation which is thus exposed for a short distance back from the scarp.

GRABENS

Introductory statement.—Smaller structural features include a large number of grabens and single faults; 24 grabens were mapped. They are expressed topographic-

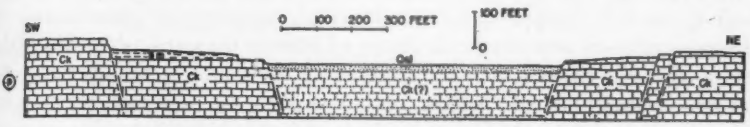


FIGURE 9.—*Structure section of Hull Canyon graben*

Letter (p) refers to position of section as indicated on Plate 2. Ck—Kaibab fm. Trm—Moenkopi fm. Qal—Alluvium.

ally as elongate valleys or trenches having rectilinear walls and with floors of Kaibab limestone often partly covered by alluvium or lava and sometimes containing patches of Moenkopi shale. Most faulting is pre-lava except locally, suggesting either a rather prolonged period of faulting or a series of distinct periods of faulting. Lithologic conditions in this region make it difficult to tell whether the grabens are original fault features or the product of erosion along ancient fault lines. The Moenkopi formation is easily washed away and could be stripped from the Kaibab limestone quickly enough to produce a "resequent graben" (graben form recreated at a lower stratigraphic level) of sharp enough appearance to suggest that its walls are recent fault scarps. Correctness of the resequent-graben interpretation is suggested in a few instances by the discovery of grabens still partly filled with Moenkopi shale which locally rises to the same or higher elevation than the limestone areas on either side of the graben.

Strata of the Alpha member of the Kaibab limestone (McKee, 1938) are so widespread and so distinctive in the Cameron-Wupatki region that they make recognition of all faults easy, through stratigraphic evidence. For most of these faults the throw is easily measured. Many small fractures, of a few feet only, are clearly revealed in the extensive bare rock outcroppings and on aerial photographs, but no attempt was made to record them. The faults mapped are all conspicuous features of the landscape.

Hull Canyon graben.—Hull Canyon, 2 miles south of the Citadel Ruin and 5 miles west of Wupatki, is 2 miles long, a quarter of a mile wide, and strikes N. 40° W. (Fig. 9). The graben has been downdropped 70 to 100 feet. The bounding faults are multiple, producing splinters along the graben sides. On both sides the Alpha member of the Kaibab limestone forms the surface rock, but Moenkopi remnants were found on splinters along the walls.

About a mile from the north end of the graben a tongue of lava has flowed over the southwest wall and across the floor. The principal fault on the southwest side of the graben passes under the flow but does not cut it, indicating that the fault antedates the lava flow. A small subsidiary fault with displacement of about 12

feet cuts this flow and produces a fresh fault scarp. Thus the principal faulting occurred before the lavas flowed out upon the region, and a second period of very minor renewed faulting followed.

Citadel graben.—The Citadel graben, near the Citadel Ruin, is 3 miles long, about a quarter of a mile wide, and trends about N. 45° E. The walls are Kaibab limestone, lava, and some Moenkopi shale. The floor, covered with alluvium, is probably underlain by Kaibab limestone.

The southwest wall from the Citadel Ruin southwestward is a rectilinear scarp about 30 feet high, capped by lavas and strongly suggestive of a fault scarp. No corresponding lavas were seen in the graben adjacent to the scarp. Near the Citadel Ruin a sink hole shows one of the faults along the southeast side of the graben. The lava is clearly faulted. Farther northeast, along the same side, the lava likewise seems to be cut by the fault. Evidence of post-lava faulting along the northwest wall also is seen in the displacement of a flow along the fault line. Although post-lava faulting has occurred on both sides, no evidence was found to prove that the entire structure was developed at that time. It was not possible to measure displacement of the limestone, which may amount to more than that developed by the small post-lava movements.

White Mesa graben.—Northward from the Citadel graben is a broad, shallow graben, 5 miles long and from a quarter of a mile to a mile wide. Its walls strike northwest in some places and northeast in others with abrupt transitions of direction, a feature unusual for grabens of this region. The walls and floor are of Kaibab limestone stripped of Moenkopi shale and scarcely dissected. The scarps are only 5 to 20 feet high but are very sharp. In places a monoclinal flexure with a narrow ridge formed from buckling at the crest replaces the fault (Pl. 4, fig. 3). Near the northern end the bounding faults are multiple and anastomosing. As no lavas lie near the structure, the date of faulting could not be determined.

In the general plan of the grabens described two directions of fracture are evident N. 30°–45° E. and N. 20°–40° W. Both trends are the same as in the Black Point section of the monocline and in the Doney and Mesa Butte faults.

S.P. graben.—The S.P. graben, extending north from a point about a mile northwest of the S.P. crater, is 8 miles long, a quarter of a mile wide, and unusually straight. The western wall is the higher, that fault having a throw of about 200 feet, as compared to 45 feet on the eastern side. Thus the plateau block west of the graben overlooks the lower eastern block. At the northern end of the graben the western fault may be traced about a mile farther north than the eastern.

The graben is now a valley floored with the Alpha member of the Kaibab limestone which also forms the plateau surfaces on either side. No Moenkopi shale was found in the graben, and it is not known if the walls are fault scarps or fault-line scarps. Lavas, which cover most of the graben floor, are not cut by the faults.

Lavas have entered the graben through its open southern end and through narrow tributary canyons in the west wall. The eastern wall, being relatively low, has been in many places overflowed by lavas. A very recent flow from the S.P. crater (Johnson, 1907; Colton, 1937) has in two places poured into the graben from the east side.

Parallel with the S.P. graben a mile west of it is a single fault with the west side downthrown. Lavas have flowed against its base. A short distance south a much smaller and more recent fault cuts the lava.

Flat Tire graben.—Midway between Mesa Butte and the S.P. Crater is a graben 2 miles long and a quarter of a mile wide striking about N. 30° E. The northwestern fault is dominant with a throw of 100 feet, as contrasted with 25 feet for the opposite

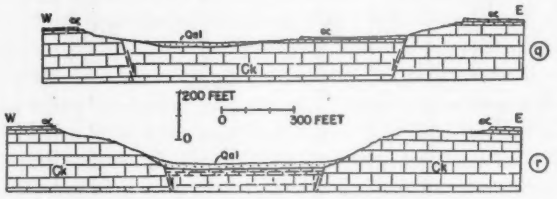


FIGURE 10.—*Structure sections of Dead Cow graben*

Circled letters refer to position of sections as indicated on Plate 2. Ck—Kaibab fm. α—Alpha member of Kaibab fm. Qal—Alluvium.

side, and is twice as long. Along the line of the graben a deep canyon has been eroded by a northwest-flowing stream. A lava flow has followed this valley north and east to enter the S.P. graben. Thus the graben clearly antedates the lava.

Dead Cow graben.—Two miles east of Mesa Butte is Dead Cow graben, a trench about 4 miles long and a quarter of a mile wide. Its northern part strikes about N. 20° W., but the southern end turns and strikes nearly north. Maximum throw of the faults exceeds 160 feet (Fig. 10). Because alluvium covers the floor at its deepest part exact figures were not obtained. Both the floor and adjacent plateaus are composed of Kaibab limestone, and no Moenkopi was seen. Near its southern end the graben has been crossed by a lava flow in a previously cut transverse valley. The lava is not faulted, and thus the graben is considered pre-lava.

Campbell Francis graben.—Two miles west of U. S. Highway 89 and a mile east of the Campbell Francis tank is a graben 4 miles long and a third of a mile wide striking nearly north. Displacements along the faults were not measured but are less than for most grabens in this area and probably do not exceed 75 feet. This graben differs from the others in that extensive areas of Moenkopi shale cover most of the floor and the area to the east. The western fault brings Moenkopi and Kaibab formations in contact along the fault. Lava flows have cut across both ends but are not faulted.

Poverty Tank graben.—A prominent graben 5 miles long, a third of a mile wide, and striking N. 25° W. extends both north and south of Poverty Tank. Throw of the faults is relatively small; the greatest measured displacement is 70 feet (Fig. 11). The graben is quite shallow and in places is still filled with Moenkopi shale, a few buttes of which rise above the surrounding plateau. As the graben lies near the crest of the Black Point section of the monocline, strata near its northern end dip 2°–3° N. E.

Just north of Poverty Tank the graben is crossed by the Tappan lava flow (Colton,

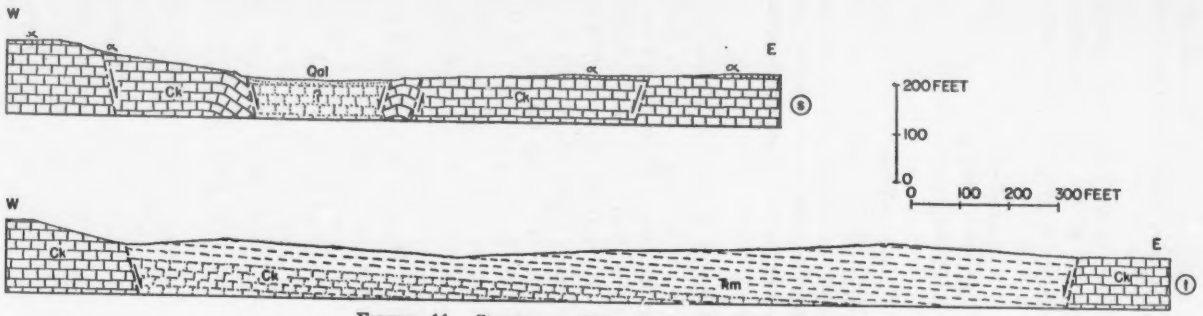


FIGURE 11.—*Structure sections of Poverty Tank graben*

Circled letters refer to position of sections as indicated on Plate 2. Ck—Kaibab fm. α—Alpha member of Kaibab fm. Trm—Moenkopi fm. Qal—Alluvium.

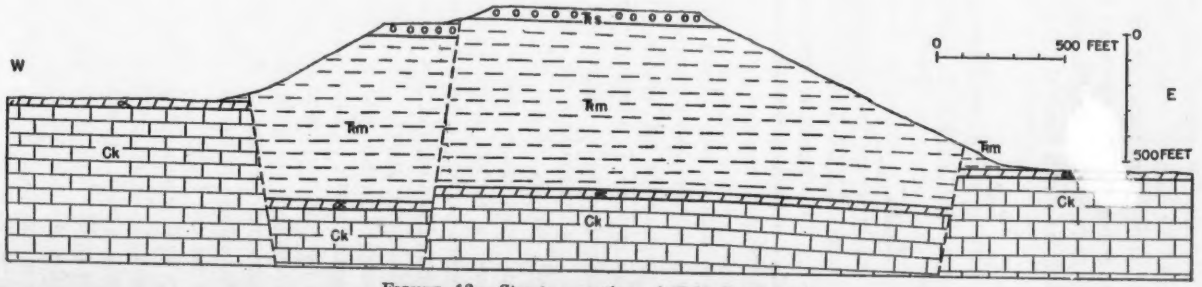


FIGURE 12.—*Structure section of Gold Butte graben*

Ck—Kaibab fm. α—Alpha member of Kaibab fm. Trm—Moenkopi fm. Trs—Shinarump fm.

1937, p. 20–22), longest known flow of the San Francisco field. The flow occupied the narrow valley of a stream which drained north into the Little Colorado River near Cameron. Examination of the flow where it crosses the faults showed no post-lava faulting. Pre-lava existence of the graben is obvious from the manner in which lava spread north a short distance along the floor.

Immediately west of the Poverty Tank graben are three similar but smaller structures. They lie in an area of Kaibab limestone. All four grabens trend about N. 25° E., oblique to the strike of the Black Point monocline.

Cameron graben.—Two and a half miles west of Cameron, just south of the Colorado River, is the Cameron graben. It has been described in detail by Reiche (1937), who has deciphered the interesting and complex Quaternary history of the immediate area. The graben is 3 miles long and about a quarter of a mile wide. Its northern two-thirds strikes N. 45° E., the southern third north. The Tappan flow, which extends across the northern end of the structure, is also faulted. This fact, together with the freshness of the scarps, 80 to 100 feet high, clearly indicates the faulting to be very recent. No evidence was found of the prior existence of this structure. Thus the Cameron graben is unique in that the entire faulting is post-lava.

Little Colorado grabens.—North and east of Coconino Point, near the Little Colorado River canyon, five grabens trend north. They vary in length from 1 to 4 miles and in width from a quarter to a third of a mile. The Kaibab, Moenkopi, and Shinarump formations are involved in the faulting. No detailed study was made of them. The displacements, though readily observable, probably do not exceed 50 feet. The Little Colorado River crosses three of the grabens but shows no significant deflection.

Gray Mountain grabens.—West of Coconino Point on the Gray Mountain Plateau seven grabens roughly parallel one another and vary in strike from north to N. 30° W. The entire area is underlain by the Kaibab limestone which dips gently southward. No detailed study was made of the structures, but the displacements vary considerably. The largest faults involved probably do not greatly exceed 100 feet in displacement.

SUMMARY

Three structural trends are followed by both major and minor structures. One trend, shown by the southern extremities of the monoclines and by the major fault lines, is N. 30°–40° E. A second, represented by part of the Black Point fold and by several grabens, is N. 25°–40° W. The third direction is north, or about midway between the first two. It is seen only in certain of the grabens. No attempt has been made to interpret the structures in regard to regional stresses which caused them. The structural part of this work is primarily descriptive.

Date of deformation of the East Kaibab monocline in this region cannot be closely ascertained. The monoclinal flexing affects Triassic and older strata and is therefore post-Triassic. Since the remaining Mesozoic series elsewhere in the region is essentially comformable, it appears safe to date the folding as post-Mesozoic.

At Black Point the monocline is unconformably overlain by horizontal Quaternary lavas. The fact that the strata throughout much of the monocline are bent, but not fractured, suggests that the deformation occurred when the region was under a thick cover of overlying conformable strata. A Laramide or early Tertiary date for the major deformation is permitted though not proven by the available facts.

GEOMORPHOLOGY

Little Colorado River.—The Little Colorado River bordering the Cameron-Wupatki region on the east and north is the master stream. Near Black Point the river is in early maturity with a narrow flood plain in weak Triassic rocks. Near Cameron, however, the stream narrows greatly and flows in an ever-deepening gorge cut in resistant Permian strata. Just north of Coconino Point, the Moenkopi and younger rocks have been removed from the Kaibab limestone, giving rise to typical development of the Marble Platform.

The Little Colorado was classified as a subsequent stream by Davis (1901, p. 165). The river follows rather closely the base of the Black Point and Coconino Point sections of the monocline, which would be the normal position for a subsequent valley developed along the strike of weak Triassic formations lying between the resistant Kaibab limestones on the west and south and the Jurassic sandstones on the north and east.

Minor drainage.—Minor streams of the region are largely resequent, subsequent, and insequent, but some examples of consequent and obsequent streams are also present. Resequents have formed wherever the Kaibab limestone is exposed and has a sensible dip. Where the dip is relatively low, as on the Gray Mountain Plateau, only the general direction of drainage is controlled, and the minor tributaries are largely insequent. Where dips are steep, as on the monoclines, streams flow directly down dip in straight, narrow canyons. Excellent examples are present in the Mesa Butte and Black Point sections. Because drainage is dominantly resequent, streams flow from the higher western and southwestern parts of the area northeast to the Little Colorado.

Subsequent streams have frequently developed along belts of weak rock faulted down into the grabens, although the fault lines themselves do not appear to have so easily controlled stream growth. Several instances were found in which streams crossed grabens with no apparent deflection of their courses. This suggests that locally stream development was more complex than already outlined, possibly including some type of superposition. Another class of subsequent streams includes those which have grown along the axis of the monocline where it has been deeply breached. Examples are present along the Coconino Point section where subsequent streams have worked deep into the fold axis, detaching the row of flatirons from the great escarpment (Pl. 3, fig. 1).

Consequent drainage is limited to the southern part of the area where lavas have flowed into pre-existing stream valleys. Streams have in several instances taken up consequent positions between the flow and valley wall, or between two flows (Colton, 1937, p. 44–45). Some obsequent streams flow against the dip along breached parts of the monocline.

Pediment remnants.—Along the base of the Black Point section between U. S. Highway 89 and Black Point are small erosional remnants of three pediment surfaces beveling the Moenkopi strata dipping 6°–20° N. E. The middle level is about 40 feet above the lowest one, the uppermost one about 140 feet above the middle one. A thin, but distinct layer of coarse, white partly rounded gravel derived from the Kaibab limestone caps the pediment remnants and contrasts sharply with the reddish Moenkopi shale beneath.

The pediments are explained by lateral stream corrasion (Johnson, 1931) during erosion of the weak Moenkopi shale. Resequent streams flowed northeast down the freshly stripped upper part of the Kaibab limestone surface, thence across the weak shales and sandstones at its base to the graded Little Colorado River. Small tributaries flowing down the monocline would thus have had opportunity to grade their courses in the weak rock. Although held in place in narrow canyons cut in the Kaibab limestone, they could swing laterally across the weak rock, planing a smooth surface across the dipping strata and leaving a thin layer of gravel. Repeated lowering of the Little Colorado may account for recutting of the pediment at successively lower levels, leaving small remnants of the earlier surfaces. There is the additional possibility that pediment recutting resulted from climatic changes or drainage modifications varying the load or volume of the pediment-cutting streams so as to permit regrading of their profiles at lower levels. There seems not to have been time for the pediments to be extended into the hard Kaibab limestone because resistance of the latter to corrasion must be enormously greater than that of the weak shale. Farther north along the monocline, 5 miles south of House Rock, the limestone has been beveled at the inner margin of otherwise similar pediment remnants, but only to a very small extent as compared with planation on the weak rocks.

Possibly the Black Point flow, which spread across beveled weak Triassic rocks of this same fold, preserves a local pediment surface. Robinson (1907, p. 112–113) cited this case as one where a small portion of a widespread peneplane of the Colorado Plateau region has been preserved under lavas. In view of the occurrence of pediments on the same monocline at lower levels, the pediment hypothesis seems the more reasonable. Thus explained, the surface underlying the Black Point lavas would not necessarily have broad cyclic significance. No examination was made of the surface on which the Black Point lavas rest. If this should be a smooth, almost plane surface, sloping 2°–3° N. E., and capped by a layer of gravels, the pediment hypothesis would be favored greatly. Even on the most advanced peneplane surfaces, resistant rocks rise above the weak ones. Moreover, on a peneplane residual soils would be expected rather than a layer of coarse stream gravels. This and other similar localities where lavas lie across monoclines or faults must be studied further before either hypothesis is validated. Data will be difficult to obtain at Black Point because a thick layer of lava talus masks the contact.

Excepting possibly the Black Point area, no evidence of any peneplane was found in the Cameron-Wupatki region. All areas of low relief are stripped limestone surfaces, constructional lava surfaces, or small pediment remnants, none of which necessarily has cyclic significance.

GRAND CANYON DIVISION

GENERAL DESCRIPTION

This division of the East Kaibab monocline passes through the eastern end of the Grand Canyon (Matthes and Evans, 1927) (Fig. 1; Pl. 5). The Canyon rim along the north side of the Nankoweap Amphitheater is the northern limit. The southern limit is placed where the Waterloo Hill and the Grandview monoclines join to produce the single monocline of the Coconino Point section.

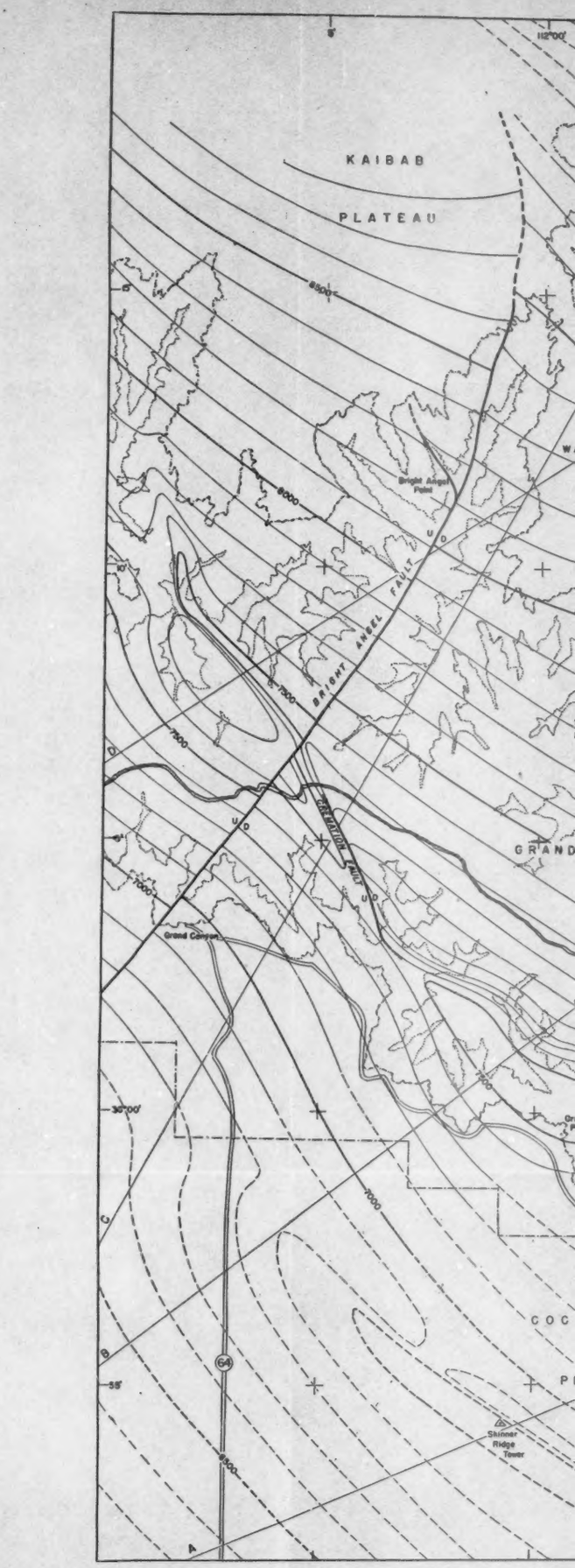
Tracing the monocline from south to north, one first sees the Waterloo Hill section, which extends north almost to Cedar Mesa, where the fold bifurcates. The western branch, or Desert View section, trends northwest to the Grand Canyon rim at Desert View, then turns north, following the canyon rim for 6 miles. The eastern branch, or Cedar Mesa section, runs north past Cedar Mesa, then turns northwest to rejoin the Desert View fold at the Grand Canyon rim. The area between the two flexures is a lozenge-shaped platform between the lower Marble Platform on the east and the higher Kaibab block on the west (Pl. 5).

The monocline turns northwest across the Grand Canyon at a point about midway between Desert View and Cape Solitude. The displacement in this section is taken up mostly by the Butte fault, which can be traced north from the Colorado River to the head of the Nankoweap Amphitheater. There the fault dies out, and the throw is in a single great monoclinal fold, clearly exposed in cross section in the Canyon walls. From this point north the monocline is in the third, or Kaibab, division.

The western branch of the monocline, formed where the Coconino Point section bifurcates in the southeast corner of the area (Pl. 5), is designated the Grandview section. It trends slightly south of west for 8 miles, then turns sharply north to intersect the Grand Canyon rim near Grandview Point. The monocline can be traced northwest in the canyon to a point several miles beyond the Colorado River.

STRUCTURE

Structure map.—Details of the monocline and adjacent areas in the eastern Grand Canyon are shown by structure contours (Pl. 5) drawn approximately on the upper surface of the Beta member of the Kaibab formation (McKee, 1938). The surface of the Kaibab and Coconino plateaus coincides closely with stratigraphic surfaces. Paleontologic evidence shows that the Beta member consistently upholds the plateau surface. This member is a massive, resistant limestone contrasting with the higher Moenkopi shales. Thus it seems reasonable to suppose that the highest elevations over the Kaibab and Coconino plateaus represent approximately the top of the Beta member. The plateau areas were contoured accordingly; the higher interstream elevations determine location of contours. The general correctness of this method is borne out by a study of the Grand Canyon Topographic Map (Mattnes and Evans, 1927) on which the canyon rim is everywhere a fairly constant distance above stratigraphic horizons in the upper Canyon walls. It is doubtful if, with the limited data available, any other horizon would be as satisfactory as the one used here. If a lower horizon were chosen, no elevations could be obtained over much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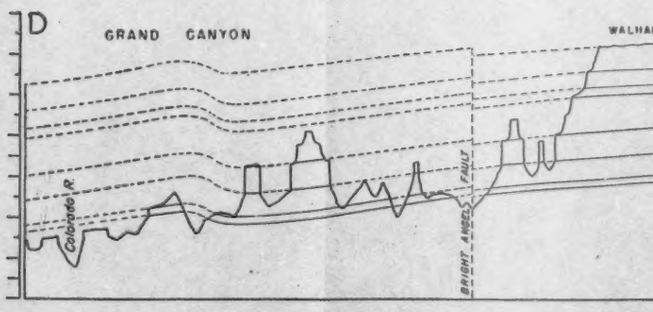
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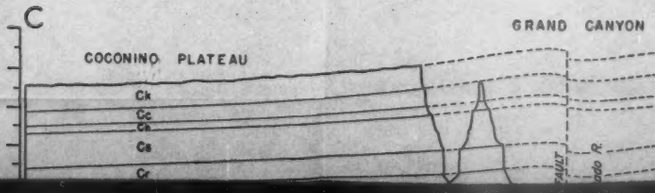
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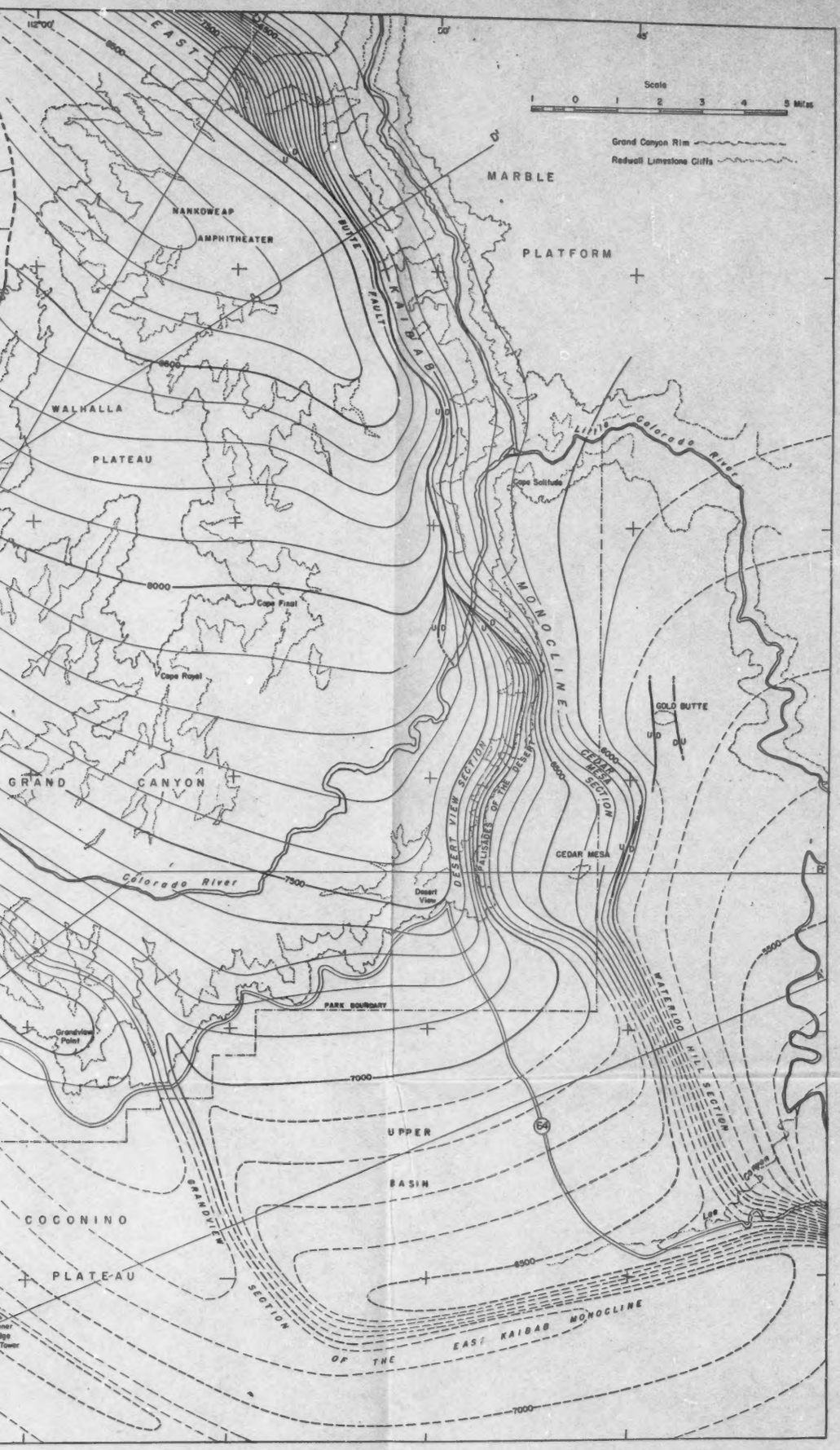
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GOLD BUTTE
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DESERT VIEW SECTION
PALISADES OF THE DESERT
CEDAR MESA SECTION
CEDAR MESA
Colorado River
Desert View
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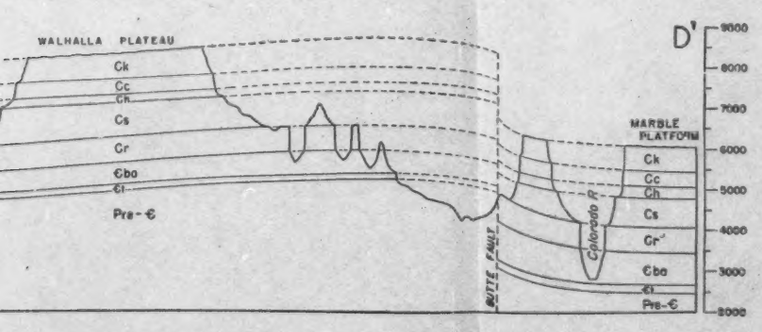
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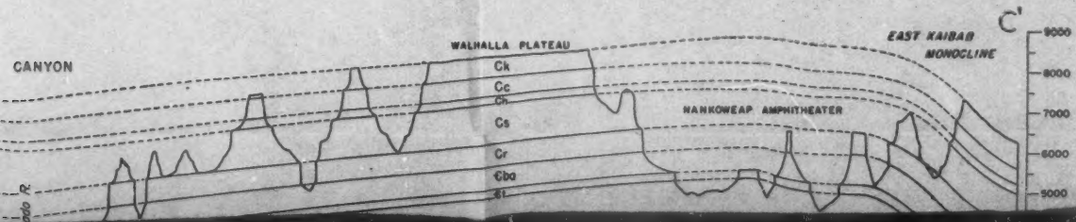
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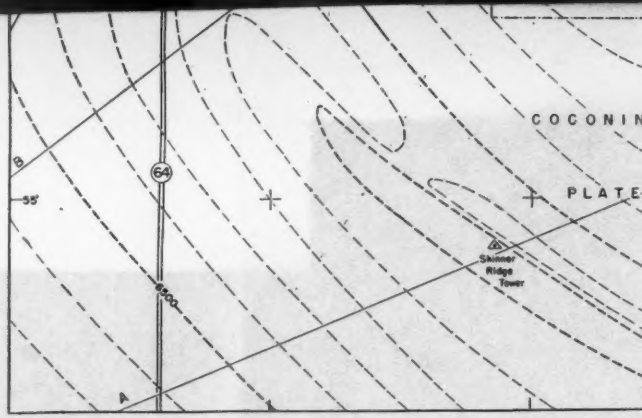
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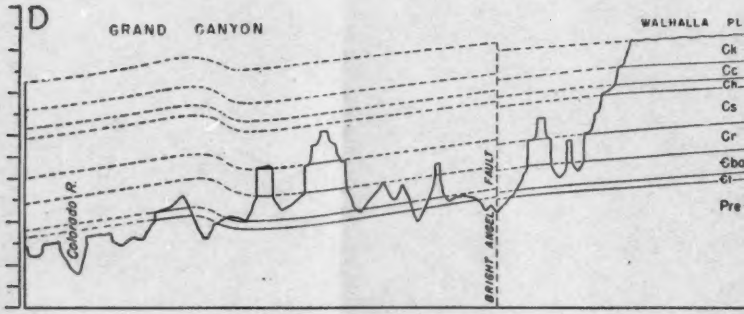
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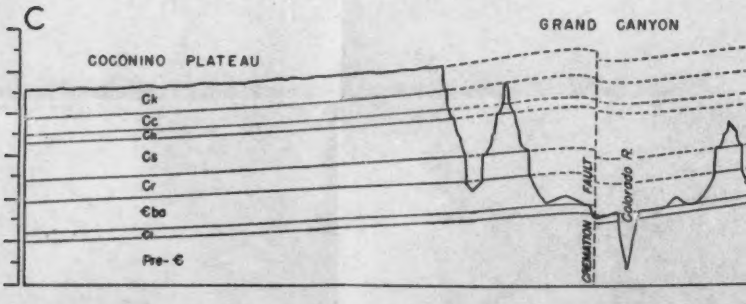
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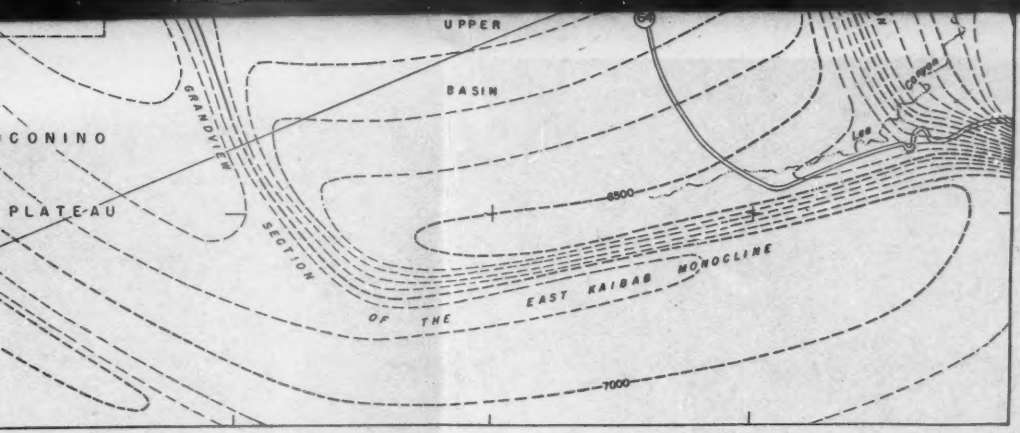
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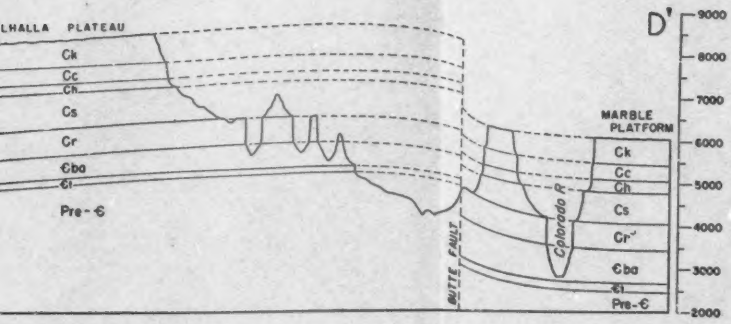
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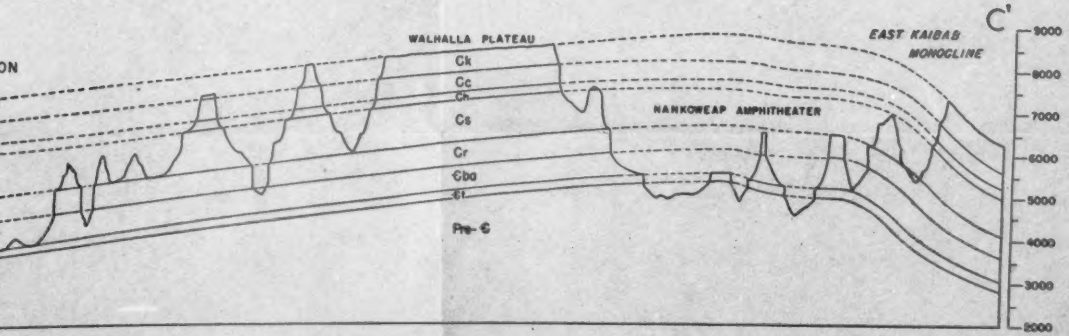
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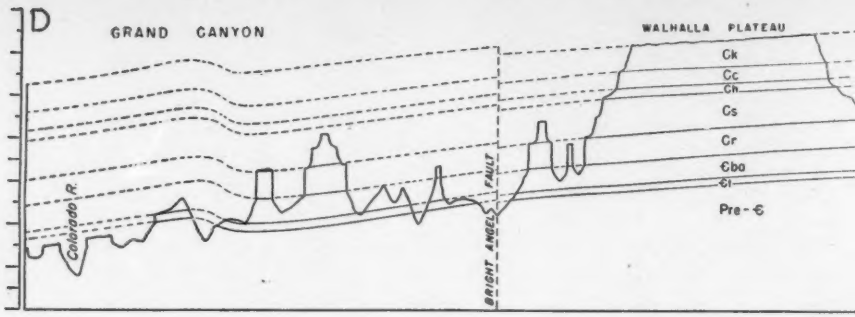
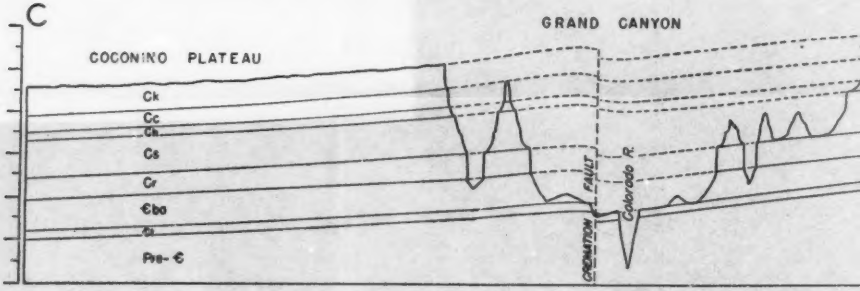
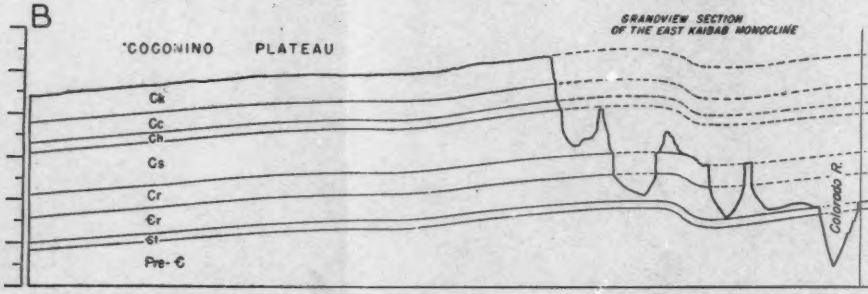
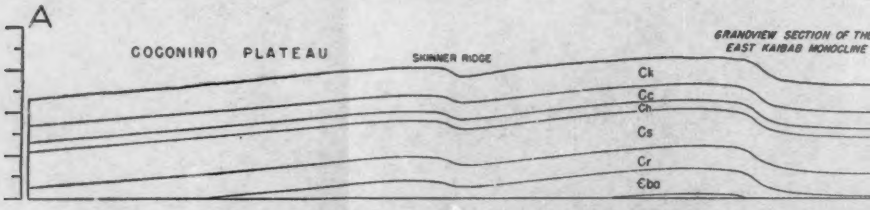


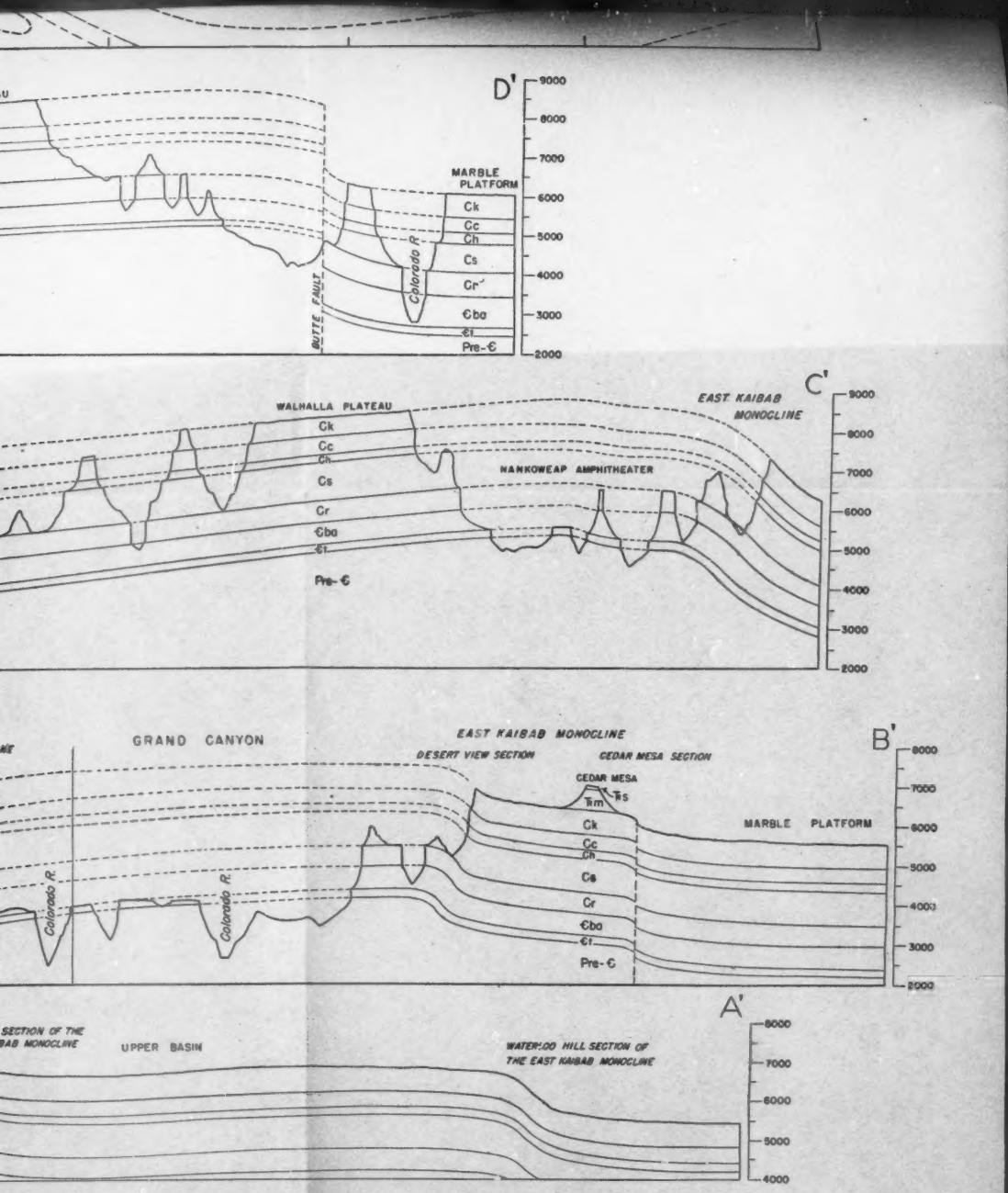






STRUCTURE MAP AND STRUCTURE SECT

Contours approximately on top of Beta member of Kaibab



URE SECTIONS OF THE EASTERN GRAND CANYON AREA

of Kaibab limestone. Contour interval 100 feet. Dashed contours unreliable.

of the plateau area, while higher horizons are almost entirely removed from the region. Within the Grand Canyon itself the bench immediately above the Redwall limestone cliff was used as a basis for determining structure. Certain easily recognizable and consistent changes in elevation of this horizon were found. Thus that part of the structure map (Pl. 5) based on the Grand Canyon Topographic Map is considered to show with fair accuracy the form, location, and approximate vertical position of the various structures but should not be relied upon for precise quantitative data. In the southern part of the map contours have been only roughly sketched from very limited data and are almost devoid of vertical control. The map is believed to show many details of structure not heretofor known or fully recognized.

Waterloo Hill section.—This section branches at nearly right angles from the west end of the Coconino Point section. The other branch continues west without interruption as the Grandview section (Pl. 3, fig. 2). Despite its sharpness, the junction of the two flexures was accomplished by bending. These structures seem to be contemporaneous. Throw of the single Coconino Point flexure is divided about equally between its two westward branches which are separated by a sloping tread or platform called the Upper (Coconino) Basin.

From its southern end, the Waterloo Hill fold extends N. 25° W. for 7 miles, then bifurcates. The throw increases from about 600 feet at its south end to slightly more than 1000 feet at its north end. The flexing is accomplished within a horizontal distance of about a mile. Maximum dips observed range from 22° to 30° and are highest at the north end. Because of a southward regional dip of about 1°, the entire monocline descends toward the south.

The section is underlain by Kaibab limestone, as are all other parts of the monocline in this region except that portion within the Grand Canyon. At the foot of the flexure most of the uppermost, or Alpha, member of the formation is present, but it has been stripped from the fold itself. A few patches were found at the summit of the fold where the beds are nearly horizontal.

Although a view down the length of the monocline shows almost a full arc of Kaibab strata, the fold is cut into by many deep canyons and is maturely dissected. Because the limestone has not been cut through entirely, there is no development of longitudinal subsequent valleys. All the canyons except one are a mile or less in length because the drainage divide lies but a short distance west of the fold summit. Lee Canyon, the exception, is a major drainage line. It flows east across the southern end of the monocline and drains the Upper Basin into the Little Colorado.

Desert View section.—Two miles south of Cedar Mesa and 3½ miles east of Desert View the Desert View section branches from the Waterloo Hill section. From this point it trends N. 65° W. and reaches the Grand Canyon rim just east of Desert View. Here it turns north and parallels the Canyon rim for 6 miles. It then turns northwest into the Grand Canyon where it is designated the Butte section. Two distinct trends are represented in the Desert View flexure—one northwest, the other slightly east of north. The throw is about 600 feet near Cedar Mesa but increases to 900 feet at Desert View, an amount which is maintained for the remainder of the distance north along the Canyon rim.

Between Desert View Point and the southeast end of the Desert View section the

monocline is similar in appearance to that portion in the Waterloo Hill section. Short streams in deep canyons flow down its slopes from a drainage divide near the summit. From Desert View Point north, however, the topographic expression of the fold is entirely different from any part so far described. The eastern wall of Grand Canyon, a great escarpment over 3000 feet high, coincides with the axis of the monocline. This cliff, named the Palisades of the Desert, faces opposite to the dip in the fold and is therefore the product of obsequent streams. It has been extended back into an earlier, east-facing resequent stream slope similar to that now present all along the fold farther south. The over-all form is that of a cuesta or low hogback. The contrast of slopes is extreme, and the drainage divide, which coincides with the Canyon rim, is being pushed rapidly eastward. Truncated shallow valleys of once longer east-flowing streams are seen in cross section along the Canyon rim. Because of the strong east dip of the strata, this part of the Canyon wall is nearly rectilinear in contrast to the intricately embayed rims elsewhere throughout the Grand Canyon. Furthermore this is one of the steepest parts of the Grand Canyon wall. At the Palisades of the Desert the Redwall cliff (2500 feet below the rim) is within a quarter of a mile horizontally of the rim. Elsewhere throughout the Grand Canyon it is at least two to three times this far. Outlying buttes and pinnacles, so characteristic of the Canyon farther west and north, are conspicuously lacking here. Just east of Desert View is a small but prominent alcove which, eaten deep into the Canyon wall, is a breach in the monocline, cut along its axis. The steep dip of the strata at this place is clearly shown in contours on the Grand Canyon Topographic Map.

Cedar Mesa section.—The eastern branch from the Waterloo Hill section, beginning 1½ miles south of Cedar Mesa, has been designated the Cedar Mesa section. It trends N. 20° E. in a nearly straight line for 3 miles, then turns sharply to N. 55° W. until it joins the Desert View section at the Canyon rim. Between the two flexures is a lozenge-shaped platform or tread in which the Kaibab strata dip gently eastward. On this platform stands Cedar Mesa, a butte of Triassic Moenkopi shales capped with thin but resistant Shinarump conglomerate (Pl. 6, fig. 1). It is one of the few remnants of a great thickness of Mesozoic strata that formerly covered much of the Grand Canyon region.

The Cedar Mesa section has relatively small displacement east of Cedar Mesa. The throw is 350 to 400 feet and is partly taken up in a fault coinciding with the line of sharpest flexing. As the monocline is traced northwest toward the Canyon rim the throw increases to about 600 feet where it joins the Desert View section. As with the monocline farther south, the upper strata of the Kaibab limestone here form the surface of the flexure and are dissected by numerous, narrow canyons.

In line with the fold axis east of Cedar Mesa, but a short distance north of the place where it turns northwest toward the Canyon, is a graben 2 miles long and half a mile wide (Fig. 12). It is similar to those of the Cameron-Wupatki region. Throw on the western fault is about 400 feet, on the eastern about 200. The graben is now largely floored with the Alpha member of the Kaibab, which also caps the upthrown areas. In the northern end of the graben is a butte of Moenkopi shale, known as Gold Butte, capped with Shinarump conglomerate. Its summit lies 400

FIGURE 1.—CEDAR MESA
A remnant of Moenkopi shale capped by Shinarump conglomerate, seen from Desert View. Axis of the East Kaibab monocline passes from right to left between observer and the Grand Canyon wall.

FIGURE 2.—GRANDVIEW SECTION OF EAST KAIBAB MONOCLINE
View southeastward across Upper Basin.

FIGURE 3.—GRANDVIEW SECTION OF MONOCLINE IN THE GRAND CANYON
Flexing of strata gives tilted appearance to Sinking Ship Butte, upper left. Monocline reappears in Redwall and Supai beds of the more distant butte in lower right.

EAST KAIBAB MONOCLINE AT THE GRAND CANYON

feet above the limestone area west of it and 600 feet above the corresponding area east of it. Its presence clearly signifies that the graben faulting is of ancient date. During stripping of the Mesozoic strata from the region the original fault trough was reversed, the downthrown strip of Shinarump conglomerate producing an elongate mesa bounded by obsequent fault-line scarps. As Gold Butte is being eroded from the graben, a trough similar to the initial graben, but on the level of the upper surface of the Kaibab limestone, is appearing. To this last form the term "resequent graben," used in connection with similar features in the Cameron-Wupatki region, seems applicable. The Gold Butte area thus provides interesting examples of the sequence of topographic forms produced during erosion of alternate resistant and weak formations along fault lines.

Many small faults, parallel with the larger structural features, were observed in this area but not mapped. A traverse east from the Cedar Mesa section toward the Little Colorado River revealed several small faults downthrown on the east side, stepping the strata downward toward the river. Straight Canyon, a deep rectilinear canyon trending due east about a mile north of Cedar Mesa, appears to be a subsequent stream developed along the line of a small fault. The common trend of small fractures is, however, nearly north.

Butte section and Butte fault.—From a point on the Grand Canyon rim 6 miles north of Desert View, the monocline runs north and northwest through the Canyon, emerging at the north wall of the Nankoweap Amphitheater. This 15-mile segment, which for most of its length coincides with the great Butte fault, is here designated the Butte section. The Butte fault and associated monocline have been described in detail by Walcott (1890; 1895). His study was made possible by construction of a special trail into the Nankoweap Amphitheater. Both because of the thoroughness of Walcott's work and the present inaccessibility of that part of the Grand Canyon, no further study of the monocline and fault inside the Canyon was made by the writers. Information supplementing Walcott's was obtained through study of the Grand Canyon Topographic Map. Interpretation of structural data from contours on this excellent map is a relatively simple operation, and the data thus obtained are considered reliable.

From the Canyon rim near Espejo Butte, 6 miles north of Desert View, one can look down into the head of Palisades Creek which has been opened up along the line of the monocline and fault. The Redwall limestone and overlying strata are folded with no trace of faulting. A short distance to the northwest the Cambrian Tapeats sandstone and older rocks are broken by the fault. This fact was noted by Walcott (1890, p. 55) and can be verified both by observation from the canyon rim and by examination of the topographic map. Half a mile southeast of the Colorado River the bench of Tapeats sandstone shows a difference of elevation of 750 feet on the two sides of Palisades Creek; most of this is throw along the fault, and the rest is taken up in monoclinal flexing. The fault may extend far southward under the Redwall formation and older rocks along the axis of the monocline. The interpretation here favored, however, is that the fault dies out near the canyon rim because the throw of the fault, where seen in the Canyon, decreases rapidly to the south.

Followed northwest across the Colorado River, the Butte fault cuts through a mesa between Carbon Canyon and Lava Canyon, bringing pre-Cambrian lava into contact with Tapeats sandstone. Since no post-Cambrian rocks are present here, it is not known whether the faulting fractured the Redwall limestone and higher strata. The structure map (Pl. 5) shows that all Paleozoic beds have been affected in the area farther north.

A mile northwest of the Colorado River the Butte fault joins a short branch fault coming in from the south. This branch dies out near the river, a mile or 2 to the south, and seems relatively insignificant in amount of throw.

Traced north the Butte fault rapidly increases in throw. For more than 10 miles it runs in generally north 1½ to 2 miles west of, and parallel to, the Colorado River. Between the fault and the river are seven buttes of Paleozoic strata for which the fault was named. Walcott recognized that the Butte fault line is very ancient and was the locus of pre-Cambrian faulting with displacement of 400 to 4000 feet. At that time the western side of the fault was downthrown. The region was later beveled, and Paleozoic and Mesozoic strata deposited. The pre-Cambrian fault does not concern us now since its topographic expression was entirely obliterated at an early date, and it has had no effect on the more recent geologic history save in the development of minor topographic features at the bottom of the Grand Canyon.

The faulting which accompanied the folding of the monocline was observed by Walcott at various points where the Redwall and younger strata in the buttes stand close to the fault line. The maximum displacement on the fault he found to be 2000 to 2200 feet near Nankoweap Mesa. At this place he estimated that in addition the beds were displaced 500–700 feet by monclinal flexing, giving a total displacement of about 2700 feet. Calculations based upon the topographic map, not available to Walcott, bear out the correctness of his total figure.

About 3 miles from the Canyon rim, on the north side of the Nankoweap Amphitheater, the Butte fault rapidly declines in throw. Three miles north of the place of greatest displacement the throw is only 500 feet. Walcott indicates the fault as dying out before the Canyon rim is reached. It is true that the fault does not cut the Permian strata of the upper Canyon wall which show only the monoclinal flexing. There is, however, the possibility, as with the southern end of the fault, that it continues in the underlying beds for a long distance northward beneath the flexed Redwall limestone.

The monocline has attained its boldest topographic development where it leaves the Grand Canyon to become the eastern slope of the Kaibab Plateau. By means of the great simple fold the upper surface of the Kaibab limestone is carried down from elevations close to 9000 feet on the Kaibab Plateau to 6000 feet on the Marble Platform. There is a splendid cross section of the monocline in the north wall of the Nankoweap Amphitheater which trends at almost right angles to the fold strike. Toward the northwestern margin of the Nankoweap Amphitheater the monocline and fault turn westward so that at the north rim the monocline trends about N. 55° W.

Grandview section and Cremation fault.—From the Coconino Point section a

westward continuation of the fold, past the point where the Waterloo Hill monocline stems from it, is noted. This western branch, the Grandview section, trends S. 77° W. for 10 miles, then turns sharply northward, running N. 25° W. to intersect the Grand Canyon rim 2 miles east of Grandview Point. The displacement of about 600 feet is accomplished within a short horizontal distance. Maximum dips range from 30° to 70°, the latter being very high as compared with most other sections of the monocline. Only along the Coconino Point section does the monocline show such a sharp flexing of the strata, a similarity which is not surprising in view of the fact that the two sections are one continuous structural line. The Grandview section is a small-scale reproduction of the Coconino Point section. In both instances erosion has breached the fold axis deeply, causing it to appear as an escarpment with occasional flatirons at the base (Pl. 6, fig. 2).

Because the plateau strata dip southward from the Grand Canyon rim about 100 feet per mile, a structural basin is produced at the foot of the Grandview fold. The land surface likewise slopes down to the trough at the base of the monocline. The topographic depression thus produced is known as the Upper (Coconino) Basin (Pl. 6, fig. 2). From the top of the monocline, the strata again slope southward about 100 feet per mile, making a structural crest, or highly asymmetrical anticline, immediately south and west of the axis of the monocline.

Six miles south of Grandview Point an additional small monocline trends parallel with the near-by Grandview fold and is also downthrown on the northeast. The flank of this monocline dips 8°. Skinner Ridge, upon which a fire tower of that name is situated, is the anticlinal structure developed by the monocline forming one limb and the southward regional dip the other.

The Grandview section may be traced inside the Grand Canyon by noting the flexings of strata in buttes and promontories. The fold passes out into the Canyon at the head of Hance Creek between the Sinking Ship and Coronado Butte (Pl. 6, fig. 3). A small fault was observed near the base of the fold with a throw of 60 feet reversing that of the monocline, which is here about 400 feet. The head of Hance Creek appears to have been opened up along this fault line.

The monocline is next seen 1½ miles northeast of Grandview Point where it is trending more to the west. Bending of the Redwall strata is clearly visible (Pl. 6, fig. 3). In the neck of Horseshoe Mesa near O'Neil Spring the monocline flexes the Redwall strata, trends N. 65° W., and has a throw of 400 feet. This locality is accessible by means of the abandoned Grandview Trail. Westward the fold reappears in the Redwall cliffs and can thus be traced to the saddle between Shoshone Point and Newton Butte. Here the displacement seems to be taken up by the Cremation fault which lies along the same line extended to the northwest.

The Cremation fault extends from the saddle between Newton Butte and Shoshone Point N. 25° W. in the valley of Cremation Creek for about 3 miles. The edge of the Tonto Platform is the farthest northwest point to which it could be followed, and no suggestion of its continuation could be seen on the north side of the Colorado River. The fault was accordingly interpreted as dying out south of the river. The fault is downthrown on its northeast side with a maximum throw of about 285 feet.

If the line of strike of the Cremation fault is extended beyond the Colorado River and Bright Angel Creek, it coincides with the canyon of Phantom Creek. A comparison of the elevations at the top of the Redwall cliffs on either side of this valley shows that those along the southwest side (Cheops Pyramid, Isis Temple) are 100 to 200 feet higher than those along the northeast side despite the fact that the regional dip of the strata is southwestward. No sharp break in cliff contours is present at the head of Phantom Creek to indicate a fault, so a fold is postulated. Confirmation of this is found a mile southwest of the Tiyo Point bench mark where the contours suggest smooth bending of the strata. On the basis of these observations a continuation of the Grandview monocline from the Colorado River to the head of Phantom Creek has been indicated (Pl. 5).

There seems to be no evidence of a further continuation of the monocline to the northwest. In any event this structure cannot be a direct continuation of the West Kaibab fault line, as suggested by Dutton (1882, p. 162–163) and Davis (1901, p. 165), because these two structures do not lie along the same line and are downthrown on opposite sides. The Grandview monocline-Cremation fault line is transected by the Bright Angel fault. Deep erosion by the Colorado River and Bright Angel Creek makes it difficult to obtain precise data at the point where the two structures cross.

Comparison of structural trends in eastern Grand Canyon and Cameron-Wupatki areas.—The monoclinal flexures and faults of the eastern Grand Canyon area follow three principal structural trends. The most prominent is a northwest strike. The East Kaibab monocline near the head of the Nankoweap Amphitheater, segments of the Desert View and Cedar Mesa sections of the same fold, and the Grandview fold and Cremation fault in the Grand Canyon strike about N. 50° W. This dominant trend is further emphasized by the axis of the broad, southeast-plunging Kaibab Plateau arch.

A second distinct structural trend, N. 25° W., is followed by parts of the Grandview and Waterloo Hill sections of the monocline and by the Butte and Cremation faults. This strike may be regarded by some as simply a variation of the more westerly strike noted in the preceding paragraph, but the angular difference seems sufficient (25°), and the segments connected by sufficiently sharp bends, to warrant its separate consideration.

A third prominent trend, N. 20°–35° E., is exhibited in the Bright Angel fault and by those parts of the East Kaibab monocline north of Desert View and east of Cedar Mesa. Still another trend, roughly east, is taken by the Grandview section of the monocline along the southern margin of the Upper (Coconino) Basin and its eastward continuation toward Coconino Point. This fourth direction is, however, relatively minor.

Within the Grand Canyon the walls of many amphitheaters, canyons, and minor recesses trend parallel with the larger structures. The northeast strike is most prominent in these erosional forms. Thus an extensive system of joints and minor faults seems to have developed with major structures of the region.

The three prominent strike directions might be expected to be similar to the three trends noted in the adjacent Cameron-Wupatki region. There a northwest strike

(N. 25°–40° W.), a north strike, and a northeast strike (N. 30°–40° E.) were found. If all three directions are rotated about 15° westward, they roughly coincide with the three principal trends in the eastern Grand Canyon area.

Regardless of whether any general agreement can be reached on the correlation of structure trends in the two areas, the general development is similar. Relatively straight segments of monoclines or faults are connected with sharp, angular bends. Anastomosing of monoclines with the junctions at large angles, and a strong parallelism of the separated parts, has produced parallelogram-shaped blocks such as the Cedar Mesa platform or the Upper (Coconino) Basin. Although it has been common practice to describe the major structures as trending north, this concept must be discarded for monocline and fault axes in the eastern Grand Canyon and Cameron-Wupatki areas. Instead the monoclines and faults proceed by a jogging series of rectilinear segments alternately oriented northwest and northeast.

GEOMORPHOLOGY

Introductory statement.—Like the Cameron-Wupatki region, the eastern end of the Grand Canyon and the immediately surrounding area may be classified as one of monoclinally flexed plateaus. In few places are the strata truly horizontal, for regional dips southward of 75–100 feet per mile prevail between the monoclines and faults. The departure from horizontality is not enough, however, to prevent classification of most of the region as a plateau. This is verified by the tendency to dendritic growth of canyons and amphitheaters in the Grand Canyon.

The development of characteristic topographic forms within the Grand Canyon has been treated by Dutton (1882, p. 140–182), Davis (1901, p. 167–187), McKee (1929; 1933), and others. In the present paper only those features related to the East Kaibab monocline and associated structures, or to the origin of the Colorado River and its tributaries, will be discussed.

Minor drainage.—The eastern Grand Canyon region as a whole is maturely dissected. Despite the fact that most of the Coconino Plateau, Kaibab Plateau, and Marble Platform surfaces coincide approximately with the top of the resistant Kaibab limestone, they are covered with an intricate system of stream valleys. Two very different scales of dissection exist side by side. The Grand Canyon, with its deep side canyons and amphitheaters, possesses enormous relief and boldness of interstream residual forms. The plateau surfaces have, by comparison, but slight relief and are devoid of the shelf-slope-scarp development characteristic of the former.

Another contrast between the plateau surfaces and the Grand Canyon interior is that on the plateaus drainage is almost entirely controlled by dip of the strata and provides one of the finest and most extensive examples of resequent drainage development. In the Grand Canyon, however, insequent and some subsequent streams dominate. Of the two types, the resequent plateau drainage is the earlier and is being encroached upon by the steep headwalls of the Canyon tributaries. Wherever the strata dip away from the Canyon rim, so that it forms a drainage divide, the rim is indented by cross sections of beheaded resequent valleys. Instances abound along the south rim of the Canyon but are best displayed along the

Palisades of the Desert where steeper dips of the strata prevail. Along the north rim of the Canyon where resequent streams flow southward over the brink into the heads of amphitheaters, hanging valleys result. The canyon rim is there a retreating fall between the steep gradient of the Canyon streams and the gentle gradient of the plateau streams.

Asymmetry of Grand Canyon.—The problem of asymmetry appears to be related to the drainage features. Between and including the Powell and Walhalla plateaus, the Colorado River lies much nearer the south rim than the north. The south rim is 2 to 4 miles from the river, and outlying promontories and buttes are not abundant. The north rim lies 6 to 10 miles from the river and is characterized by deep embayments, with long projecting peninsulas between, and numerous outlying buttes and mesas. It might be supposed that the Colorado River has slipped laterally down the dip of the strata which is to the south and southwest. Though this phenomenon doubtless operates where dips are strong, it is unlikely that a dip of less than 2° would have a noticeable effect. Moreover, the presence of buttes and mesas close to the river on both sides, and about equidistant from it, shows clearly that widening of the Canyon is primarily the product of weathering and tributary-stream erosion.

Asymmetry of the Grand Canyon has been explained by Noble (1914, p. 26–27) and Matthes (1932). Drainage from a large area of the Kaibab Plateau, both surficial and subterranean, flows southward over the rim or from springs, adding its volume to tributary canyons sapping at the heads of the great amphitheaters. The increased volume makes possible more rapid recession of the canyon heads. On the south rim plateau streams and presumably much underground water flow away from the rim, which is a drainage divide. Here the canyon tributaries receive only the water which falls within the limits of each.

A second factor contributing to asymmetry is that of elevation (Noble, 1914, p. 26–27). The north rim with elevations of 8000 feet and more, increasing northward in the Kaibab Plateau, contrasts with the south rim, only slightly over 7000 feet and diminishing southward in the Coconino Plateau. Inasmuch as rainfall increases considerably with increase in elevation, more water must be available to streams pushing back the north rim, thus causing that wall to retreat more rapidly. Average annual precipitation on the north rim at an elevation of 8200 feet is 26 inches; on the south rim at 7000 feet, 18 inches.

A third factor, closely related to the second, is that, because the strata are lifted higher along the north rim, the weak formations there were bared to erosion before corresponding beds in the south wall. This has probably contributed to more rapid retreat of the cliffs northward into the Kaibab Plateau. The same principle attends the breaching of anticlines and domes to produce covelike valleys. The highest portion of the structure is breached first, while recession of the resistant layers progresses down the pitch of the fold.

A fourth factor, stated by Matthes (1932) is that minor fault lines which facilitate the growth of canyons are numerous north of the Colorado River but die out to the south in the Coconino Plateau. Subsequent-stream development along the Bright Angel fault line has resulted in unusually deep incision of the north rim, but the

comparatively shallow notch in the south rim where the same fault transects it suggests that factors other than greater ease of erosion along fault lines may be of primary importance.

Origin of Colorado River.—The origin of the Colorado River through the eastern part of the Grand Canyon has been speculated about since the first geological explorations of the region. The feature requiring special attention is that the river passes from the relatively low Marble Platform area westward through the Kaibab arch which resembles a great anticlinal barrier in the path of the river. This portion of the river is critical for testing hypotheses of the origin of the Colorado River.

Powell (1875, p. 198) regarded the Colorado River as antecedent to the Kaibab arch, which grew athwart the river's course. He saw only a discordance of the river and the barrier. Dutton (1882) concurred with Powell, but Walcott (1890, p. 60–61) noted two facts conflicting with the earlier interpretation. He reasoned that a great thickness of overlying strata must have been present at the time of flexing of the East Kaibab monocline, thus requiring the fold to be much older than previously supposed. He nevertheless considered the antecedent explanation as the most likely, although the river at that time was, he supposed, flowing in the overlying Mesozoic rocks. A second important fact which he recognized is that the Colorado River in the eastern end of the Grand Canyon is closely related to rock structure, contrary to Powell's belief. Walcott considered this part of the stream to have been considerably modified by adjustments to the structure after the Kaibab arch rose.

According to Davis (1901, p. 158–167), drainage was initially directed eastward down the steps of the monoclinally flexed blocks. With dropping down of the Basin Range province to the west, and downfaulting of plateau blocks west of the Kaibab area, a drainage reversal was initiated. Consequent streams flowed down over the block steps, westward from the Kaibab arch, which now had become the highest line of elevation of the Grand Canyon region. By the same deformation, drainage even far east and northeast of the Kaibab Plateau turned and flowed westward. Davis seems to have envisioned sufficient tilting of the land surface to cause streams to reverse their direction of flow and to become consequent on the reversed slopes of their own valleys. The lowest place available for drainage westward across the Kaibab barrier was a narrow lowland strip of Moenkopi shale preserved in the trough between the Kaibab and Coconino plateaus, at the base of the Grandview section of the East Kaibab monocline (Pl. 5). The Little Colorado and that part of the Colorado River between the Grand Canyon and Lees Ferry were, prior to faulting, subsequents in an elongate subsequent lowland between the Kaibab-Coconino arch on the west and high Jurassic cliffs on the east. With drainage reversal, these streams joined and flowed westward across the Kaibab uplift in the transverse structural and topographic depression described above.

Davis had only a general conception of the local structures with which to work, and no accurate picture of relative elevations. Study of the structure map (Pl. 5) shows that the conditions Davis postulated could not have existed simultaneously. Although the depression at the foot of the Grandview monocline does cut diagonally across the Kaibab arch as a sag between the Kaibab and Coconino plateaus, this

transverse trough is not deep enough to have had Moenkopi shale preserved in its floor continuous with a subsequent lowland of the same rock running east of the Kaibab arch at the base of the East Kaibab monocline. We are thus forced to discard the details of Davis' hypothesis for this locality. Furthermore, the idea of drainage reversal as Davis pictured it seems unacceptable. The degree of tilting required for reversal seems greater than could have occurred over a large region. Nevertheless, Davis should be credited with associating the Colorado River in the eastern end of the Grand Canyon with structural features and with suggesting its development in subsequent Triassic lowlands between the Kaibab arch and Jurassic cliffs.

Discussion of the origin of the Colorado River is here limited to the relationships of the river to geological features in the eastern end of the Grand Canyon, where it crosses the East Kaibab monocline. Detailed studies of the remaining parts of the river's course should be completed before a regional analysis is undertaken.

Of the various possible origins for the Colorado River in this locality—antecedent, consequent, superposed, or subsequent—the last named appears the most satisfactory. In general plan, the Colorado River makes a great bend southward around the pitching nose of the Kaibab arch (Pl. 5). This suggests that, as the Mesozoic strata were being stripped from the region, the river may have occupied a subsequent lowland belt of weak Moenkopi and Chinle shales between the plunging nose of the Kaibab limestone arch and the encircling north-facing cliffs of Jurassic sandstones. A geological sketch map restoring the areal geology to test this hypothesis is shown (Fig. 13 A). The floor of the subsequent valley would have had to be approximately at the present 8000-foot level in order to locate the weak-rock area where the river now flows. If we take 1000 feet as the thickness of the combined Moenkopi and weak lower half of the Chinle formations, which we are assuming to have underlain the subsequent lowland, the contact between these strata and the overlying upper Chinle and Jurassic sandstones would have to be drawn along the present 7000-foot structure contour. A cliff line has, accordingly, been placed approximately along this line (Fig. 13 A). The cliff may have faced north and west and bounded the weak Triassic and lower Jurassic valley on the south and east. The subsequent valley would have been several miles wide and would have coincided approximately with the present Grand Canyon. North from the locality of Desert View, the weak-rock lowland would have narrowed rapidly as it turned to run along the Butte fault line. There it must have been in a narrow canyon whose western wall was the resequent fault-line scarp of the Butte fault formed of Kaibab limestone. The eastern wall would have been made up of Jurassic sandstones of the Glen Canyon group. Northward from the Grand Canyon, where the Butte fault gives way to the East Kaibab monocline, the subsequent valley would have widened somewhat but would still have been much narrower than the broad east-west portion.

It is necessary to find some reason for the ancestral Colorado becoming entrenched in the Kaibab arch rather than shifting down dip as the subsequent lowland migrated southward and eastward. Perhaps at this stage the river was rejuvenated, or made conquests in its upper portion, either of which may have caused it to become incised.

It is also possible that hills of Kaibab limestone began to appear through the Moenkopi shales south of the river course. These would be parts of the anticlinal axis at the crest of the Grandview section (Fig. 13 A; Pl. 5). Thus, north of Grandview Point, the river would have been caught in a shallow syncline, with higher Kaibab limestone masses both to the north and south. The river crosses the Grandview

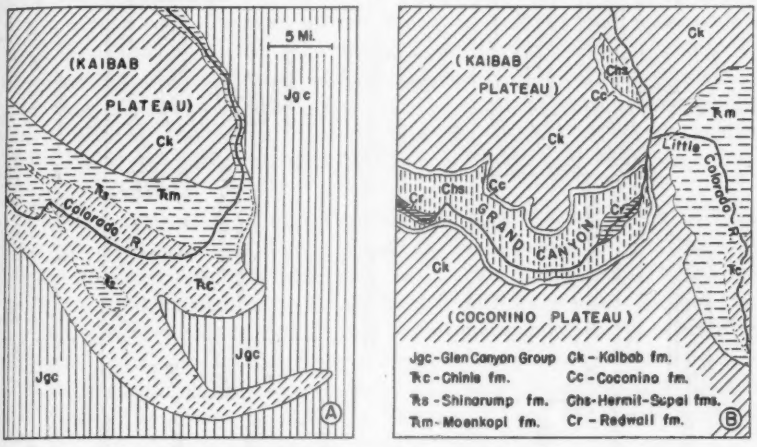


FIGURE 13.—*Geologic sketch maps of eastern Grand Canyon area*
Showing early stages in development of the Colorado and Little Colorado rivers and the Grand Canyon.

monocline-Cremation fault axis obliquely at the place where the Bright Angel fault intersects it. Structurally there is produced a notch or sag in the crest of the Grandview flexure (Pl. 5), a feature which may have determined local details of the Colorado course as it was being superposed from the weak rock onto the hard limestones. Under the general hypothesis outlined above, adjustment of the river to minor structures would not necessarily be perfect as there would have been included the element of chance superposition from weak rocks in which structures had little or no control on drainage.

The Little Colorado would have developed later in the stripping of Mesozoic strata from the region (Fig. 13 B). After the Jurassic cliffs had retreated southward and eastward, the cliff line would have lain east of the Kaibab and Coconino plateaus, somewhere along the Marble Platform. The Little Colorado may have grown southward along the weak Triassic shale belt between these cliffs and the East Kaibab monocline. Its present position, following the base of the Waterloo Hill, Coconino Point, and Black Point sections, strongly suggests this origin.

Whether this hypothesis can be fitted into a more general scheme of geomorphic evolution cannot yet be stated. The origin of the small segment of the Colorado River as outlined here does not exclude a superposed or antecedent origin for the river as a whole. If the river had in either manner come to lie across the Kaibab fold it might have, by down-dip shifting or a series of autocaptures, developed the

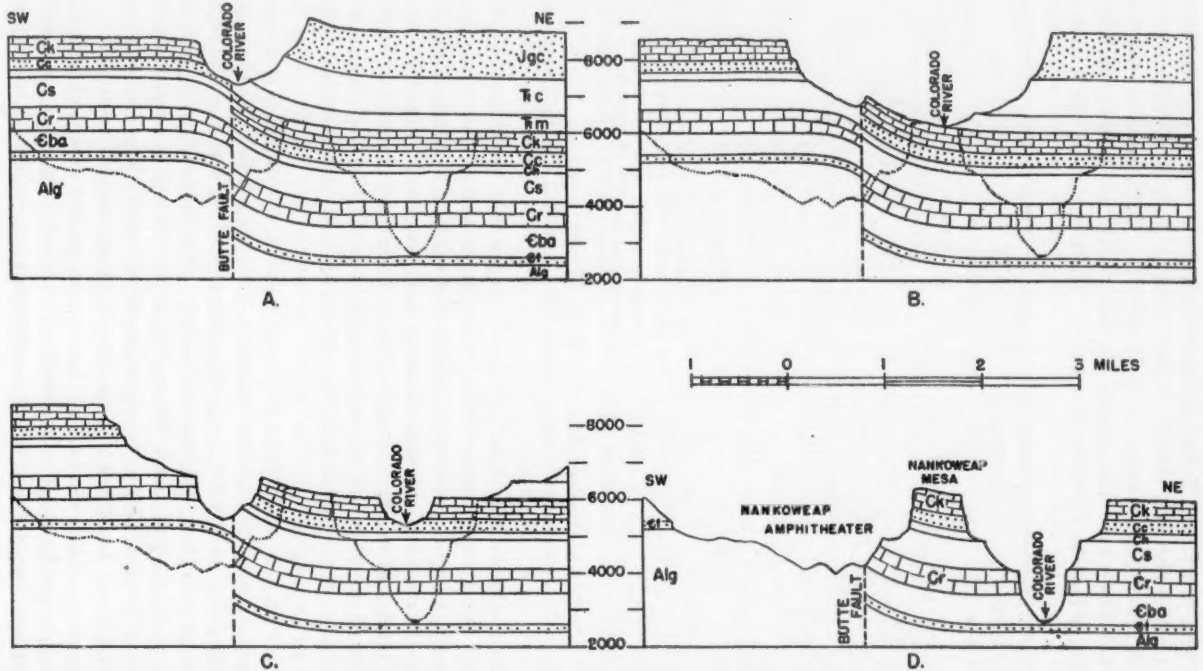


FIGURE 14.—*Restored structure sections across Butte fault and East Kaibab monocline*

Showing supposed manner of development of eastern end of the Grand Canyon. Alg—Algonkian. Ct—Tapeats fm. Cba—Bright Angel fm. Cr—Redwall fm. Cs—Supai fm. Ch—Hermit fm. Cc—Coconino fm. Ck—Kaibab fm. Trm—Moenkopi fm. Trc—Chinle fm. Jgc—Glen Canyon group.

great southward bend around the nose of the anticline. Such a history would, however, be preliminary to the series of events outlined in the preceding paragraphs.

Colorado River and Butte fault.—The interesting relationships between the Butte fault line and the Colorado River were discussed by Walcott (1890, p. 60–62). His explanation fits closely into the general hypothesis of river origin stated by the writers. The river does not now follow the Butte fault line but instead parallels it 1 to 2 miles east. Between the two lines is a chain of seven buttes made up of the Redwall limestone plus varying thicknesses of the overlying Permian strata. Three are capped by patches of Kaibab limestone. The buttes are truncated along their western sides by the obsequent fault-line scarp, on their eastern sides by the canyon walls of the Colorado River. They are separated from one another by water gaps of streams flowing eastward from the Nankoweap and Kwagunt valleys into the Colorado River.

Walcott's explanation of how the Colorado River became separated from the fault line is illustrated by Figure 14. In A, the Colorado River is a subsequent stream occupying the Butte fault line. In B, the river has cut through the weak Moenkopi and Chinle strata to the resistant Kaibab formation. Because the beds are upturned at the fault, the river has shifted eastward down the relatively steep dip slope. In C, the Colorado River has begun to incise the limestone strata which at that distance from the fault are nearly horizontal. A second subsequent valley has begun to be opened up along the Butte fault line. When the Bright Angel shales were reached, this valley was rapidly widened, and the Nankoweap Amphitheater began to develop. Buttes were being formed between the fault line and the river. In D (present conditions), relief has been greatly increased and the pre-Cambrian Algonkian rocks widely exposed in the Nankoweap Amphitheater. Rapid westward recession of the amphitheater walls has been facilitated by the weakness of the pre-Cambrian rocks. While other hypotheses or variations of the one offered are possible, Walcott's hypothesis has been utilized here because it fits closely with the general scheme of river development outlined for the eastern part of the Grand Canyon.

The Nankoweap Amphitheater has been excavated along the axis of the Kaibab arch (Pl. 5, section DD′) and is an anticlinal cove valley reflecting the trend of the larger structure. In the future this amphitheater will be extended farther northwest where it will merge with the Pagump Valley, a breached portion of the East Kaibab monocline farther north.

KAIBAB PLATEAU DIVISION

GENERAL DESCRIPTION

The Kaibab Plateau division extends from the Grand Canyon on the south to the abandoned settlement of Paria 17 miles north of the Arizona-Utah boundary (Fig. 1; Pl. 7). This division forms the east side of the Kaibab Plateau where strata of the Kaibab Plateau are flexed down to the level of the Marble Platform-Paria Plateau block.

Although in this division the monocline is a relatively simple structure, there

are certain distinctive changes in its form or topographic expression which make desirable its subdivision into the southernmost or Pagump Valley section, the central or Kane Canyon section, and the northernmost or House Rock section.

The structure-contour map (Pl. 7) was compiled from sources which differ greatly in accuracy. The new U. S. Geological Survey topographic map of the Jacob Lake quadrangle provided detailed and accurate data for mapping the Kane Canyon section. Information regarding the monocline between the Arizona-Utah line and Paria was taken from a small-scale structure map of the Kaiparowits region by Gregory and Moore (1931). The remainder of the map is based on old topographic maps of the Powell Survey on the scale of 1:250,000 supplemented by field data. Structure contours are drawn on the approximate top of the Kaibab formation.

STRUCTURE AND GEOMORPHOLOGY

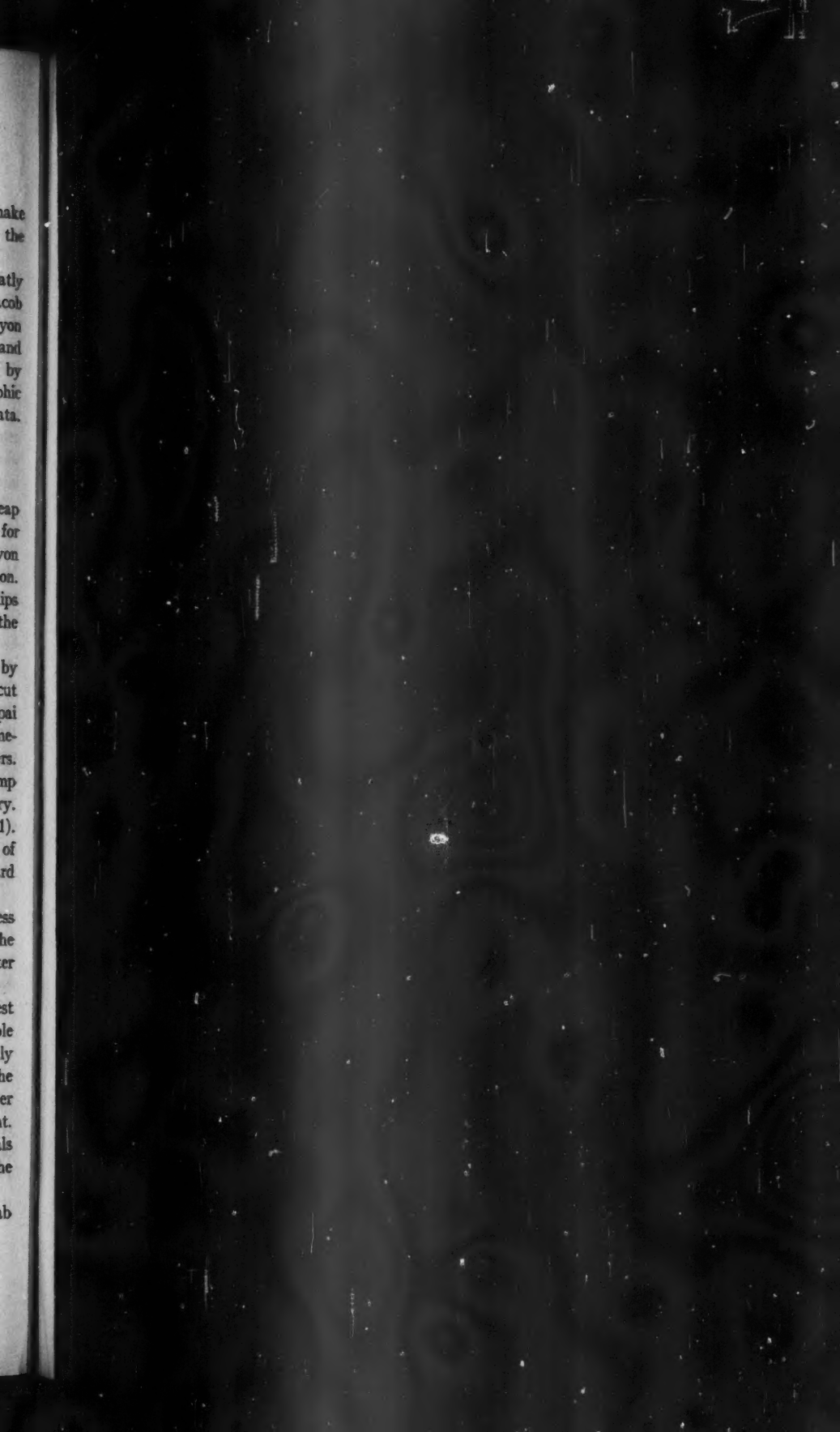
Pagump Valley section.—This section extends north from the head of Nankoweap Amphitheater and is a continuation of the Butte section. It strikes N. 55° W. for 4 miles but then turns sharply northward. Ten miles north of the Grand Canyon the monocline bifurcates. This point is taken as the north end of the section. Throw on the monocline is about 2500 feet along the section, while maximum dips range from 23°–27°. The monocline is a single, simple flexure but toward the northern end begins to show the double form.

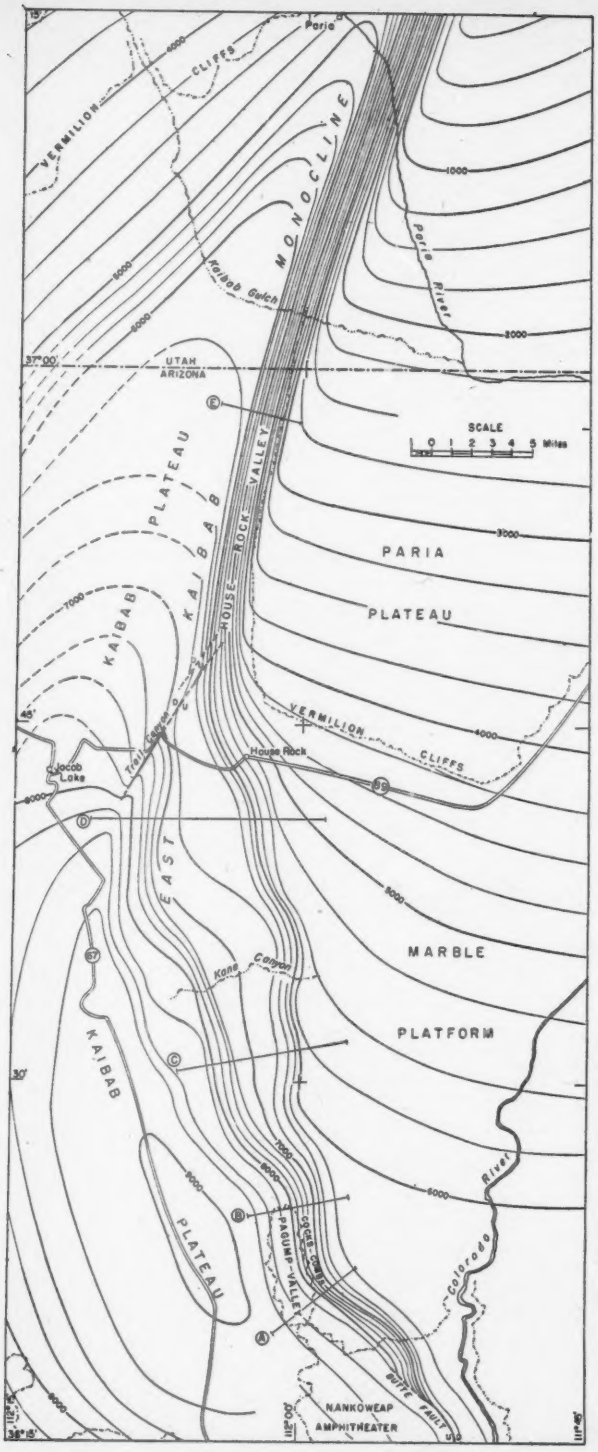
Throughout most of this section the monocline has been deeply breached by streams. Pagump Valley, a deep covelike depression 6 miles long, has been cut into the flank of the fold (Pl. 8, fig. 1). In the valley floor a large area of the Supai formation is exposed. Dutton's (1882) geologic map shows that the Redwall limestone is exposed in the valley bottom, but this formation was not seen by the writers. A splendid panoramic sketch by W. H. Holmes (Dutton, 1882) of the Pagump Valley viewed from the Kaibab Plateau is recommended for details of the scenery. A series of flatirons, the Cocks Combs, bound the valley on the east (Pl. 8, fig. 1). These are made of dipping Kaibab and Toroweap limestone beds near the base of the monocline. Pagump Valley is drained by several canyons which flow eastward across the Marble Platform into the Colorado River.

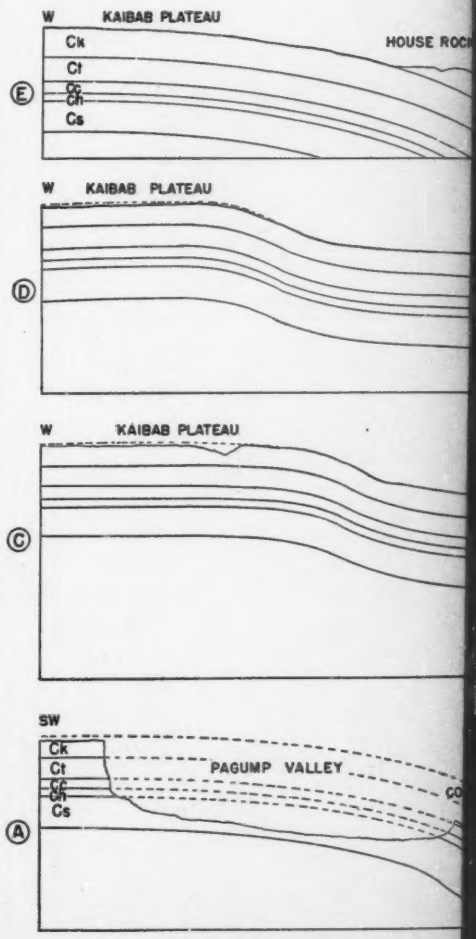
Between Pagump Valley and the Grand Canyon the axis of the monocline is less deeply eroded. In time the entire Pagump Valley section will become a part of the Grand Canyon through slow northward growth of the Nankoweap Amphitheater along the monocline axis.

The Kaibab Plateau is highest, both structurally and topographically, just west of the Pagump Valley section, where it rises to more than 9000 feet. The Marble Platform along the monocline base is also highest in this area. Its surface is slightly over 6000 feet in elevation but slopes northward, with increasing dip toward the Kaiparowits region, and southward in the direction of the Little Colorado River valley. A broad east-trending transverse anticlinal arch thus seems to be present. A structure-contour map of the Colorado Plateaus region by Darton (1925) reveals extensions of this axis westward across the Kanab Platform and eastward into the Navajo Country.

Kane Canyon section.—About 4 miles north of Pagump Valley the East Kaibab

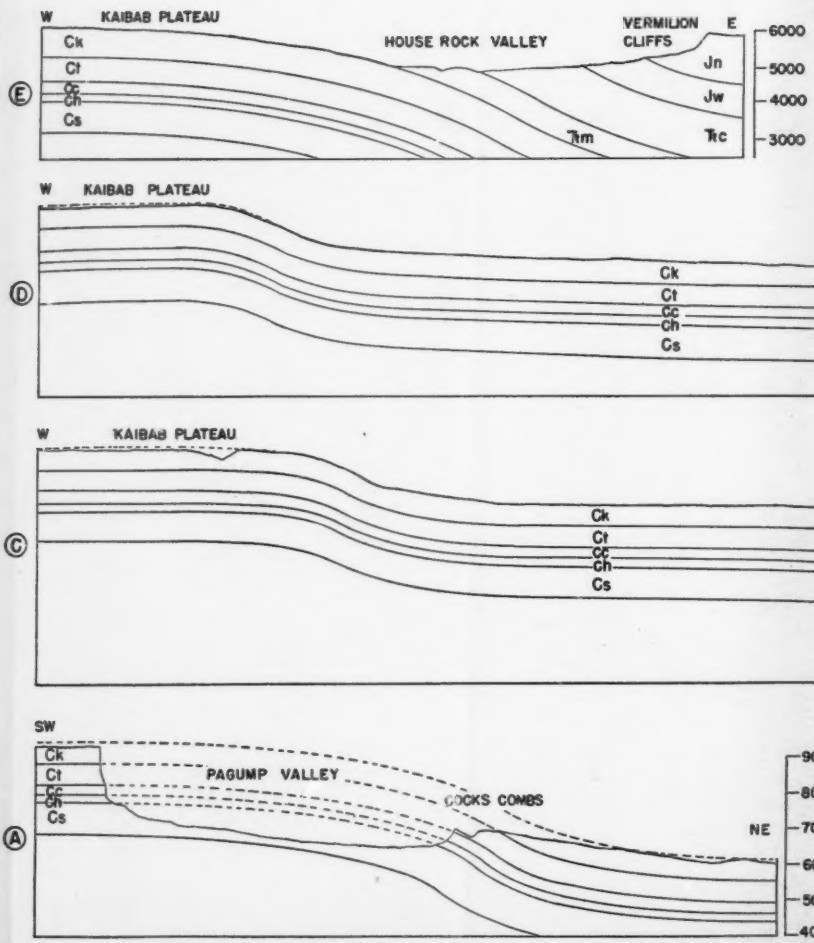






STRUCTURE MAP AND STRUCT

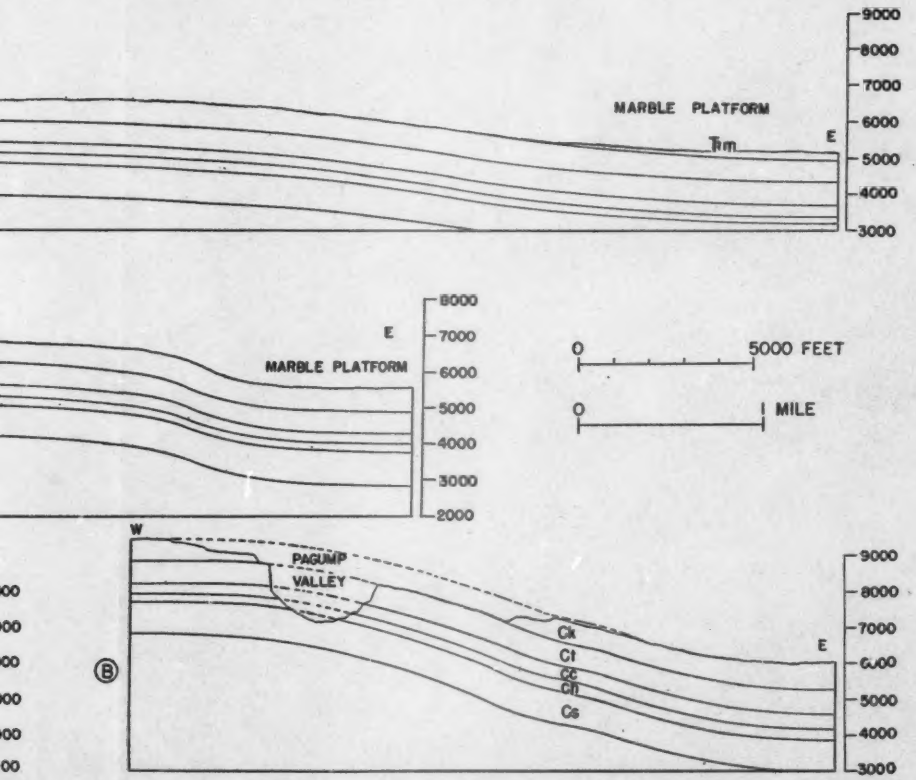
Contours approximately on top



STRUCTURE MAP AND STRUCTURE SECTIONS OF KAIBAB PLATEAU DIVI

Contours approximately on top of the Kaibab formation. Contour interval 250 feet

Jn - Navajo ss. Trm - Moenkopi sh. Cc - Coconino ss.
Jw - Wingate ss. Ck - Kaibab ls. Ch - Hermit sh.
Trc - Chinle sh,ss. Ct - Toroweap ls. Cs - Supai sh,ss.



EAU DIVISION
l 250 feet.

monocline divides into two monoclines with a platform or tread between. The flexures rejoin at the south end of House Rock Valley, forming a lenslike platform between them about 5 miles wide near the middle. This part of the monocline is designated the Kane Canyon section. Above the two principal flexures already mentioned there is a third one of minor importance. It is not, however, large enough to distract from the double appearance of the monocline.

This section trends a few degrees west of north, but there are numerous irregularities of strike, and the flexures tend to alternately bend more westerly and more easterly, following each trend for 4 or 5 miles. Total throw of the compound monocline, from the summit of the Kaibab Plateau to the Marble Platform level, is about 3000 feet. Each major flexure has a throw of 1000 to 1200 feet; the third or uppermost one, 300 to 500 feet. Maximum dips range from 9°–14°, making the flexures appear broad and gently curving (Pl. 8, fig. 2).

Kaibab limestone surfaces this section, but some thin remnants of the Moenkopi shales are present. This weaker rock was not observed near the upper part of the monocline, but small patches remain on the platform between the two major flexures. Two miles south of U. S. Highway 89 a patch of Moenkopi shale about half a mile wide seems to have been downfaulted, and at the monocline base on the Marble Platform is another rather extensive area.

Near the northern end of the section, between House Rock and Jacob Lake, the monocline is cut obliquely by a normal fault which strikes N. 30° E. and is downthrown about 200 feet on the northeast. A stream deepening the fault line produced Trail Canyon.

Geomorphology of this section has been described elsewhere (Strahler, 1942). Although the present surface of the monocline coincides closely with the upper surface of the Kaibab limestone, it is deeply scored by narrow canyons. Canyons in the upper two flexures are numerous and closely spaced, and the surface is therefore maturely dissected. Upon reaching the broad structural terrace these streams join to form a few large streams flowing northeast to the brow of the lower fold. In contrast to the upper two flexures, the lowest one is crossed by fewer streams, and these have incised themselves more deeply. Whereas on the upper folds interstream areas have almost entirely disappeared, on the lowest fold wide belts of stripped limestone surface separate the deep canyons.

Without appealing to erosion cycles to explain this contrast, an explanation may be found. One possibly adequate factor is the marked difference in precipitation on the two flexures. A rich growth of forest on the upper fold indicates that it receives much more water than the lower fold. More rapid dissection of interstream divides on the upper fold would thus be expectable. On the lower fold corresponding dissection would be much slower, but at the same time the larger streams crossing it would cut deep canyons because their headwaters extend up to the more humid Kaibab Plateau. A second factor is that, during regional denudation, the upper parts of the present fold surface would have been exposed longest, and stripping of the weak Triassic cover would progress downward.

Dissected pediment remnants, first described by Johnson (1909, p. 146–147) and found from near Kane Canyon northward into House Rock Valley, are best

exhibited near the mouth of Rock Canyon 5 miles south of House Rock where long, flat-topped, tonguelike remnants extend eastward from the monocline base (Pl. 8, fig. 2). There are two distinct pediment levels; the higher stands 150 feet above the lower, which in turn stands 90 feet above the present stream-graded land surface between the remnants. The pediment remnants consist of Moenkopi shale, dipping 6°–10° E., perfectly beveled by the two planes which slope 1½°–4° E. A layer of coarse, poorly sorted Kaibab and Toroweap limestone gravels caps the rock planes. The white-gravel layer contrasts with the red-brown Moenkopi shale and emphasizes the unconformity. Two and a half miles south of House Rock the pediment bevels not only the Moenkopi shales, but also the Kaibab limestone. Here the rock plane includes a strip of Kaibab limestone a third of a mile wide. Despite the great difference in resistance of the limestone and shale, the pediment surface has planed across both rocks impartially.

The present stream-graded land surface slopes eastward at about the same rate as the pediment surface. Several miles eastward on the Marble Platform it is dissected, revealing a cover of Kaibab and Toroweap gravels resting on Moenkopi shales and indicating that it, too, is a pediment surface still being cut near the monocline base, but being regraded farther out.

The Jacob Lake topographic map shows that the pediment remnants near the mouth of Rock Canyon are part of a rock fan having its apex where the stream debouches from a constricted canyon mouth at the base of the monocline. The fan has a radius of 2 to 3 miles, and the remnants slope northeast, east, and southeast from the apex. Pediment remnants elsewhere along the monocline are probably also small rock fans with apices at the mouths of canyons.

About half a mile north of House Rock dissected remnants of a single pediment surface slope from the monocline base eastward to House Rock Wash. From Kane Canyon south along the base of the monocline a graded alluvial surface slopes eastward and is scored by shallow stream channels. As no deep exposures were seen, it was not possible to determine if a pediment surface underlies this alluvial material.

Origin of the pediments through lateral stream corrasion is indicated by the fan form of the Rock Canyon surfaces and by the cover of stream gravels. Various parts of this area show stages in the dissection and regrading of pediment surfaces. The pediments should not be interpreted as peneplane remnants, but only as local phenomena.

House Rock section.—From House Rock to Paria, 36 miles, the monocline is a single flexure having much the same geological and topographical characteristics throughout its length. It differs from any sections of the monocline thus far described in that the Vermilion Cliffs of Jurassic sandstone lie close to the base of the fold. West of the monocline is the Kaibab Plateau which slopes northward. The monocline here as elsewhere bends the Kaibab limestone downward smoothly on the east margin of the plateau. Near the base is House Rock Valley, longitudinal, subsequent, and developed in the upturned Triassic Moenkopi and Chinle shales (Pl. 9, fig. 1). The east wall of House Rock Valley is formed by the Vermilion Cliffs bordering the Paria Plateau whose surface is formed by Jurassic sandstones. By a fortuituous combination of thickness of strata and throw of monocline, the

FIGURE 1.—PAGUMP VALLEY AND THE COCKS COMBS

East Kaibab monocline viewed from South Canyon Ranger Station. Longitudinal subsequent valley breaching axis of monocline is Pagump Valley. The Cocks Combs, a row of flatirons, formed of dipping limestone strata at base of monocline. Kaibab Plateau on skyline. The prominent white cliff in the escarpment is the Coconino sandstone.

FIGURE 2.—PEDIMENT REMNANTS AT BASE OF EAST KAIBAB MONOCLINE

Near the mouth of Rock Canyon, south of House Rock. In background is lower fold of the compound monocline deeply incised by canyons.

EAST KAIBAB MONOCLINE NORTH OF GRAND CANYON

FIGURE 1.—VIEW NORTHWARD INTO HOUSE ROCK VALLEY
Showing East Kaibab monocline on left and Vermilion Cliffs of the Paria Plateau on right.

FIGURE 2.—PARIA DIVISION OF EAST KAIBAB MONOCLINE
View southward along monocline from The Gut. Kaiparowits Plateau on left, hogback ridge of Cretaceous Kaiparowits sandstone on right.

Paria and Kaibab plateaus on either side of the valley are at approximately the same level. Hence the monocline is represented topographically not by a slope between a higher and lower plateau level, but by a valley between accordant plateau surfaces.

The House Rock section strikes almost due north at its southern end but curves progressively more eastward as it approaches the Utah border. At Paria it strikes about N. 20° E. Displacement on the monocline varies from 3000 feet near House Rock to an estimated 4000 feet at Paria. Both the Kaibab and Paria plateaus dip northward, but the dip of the Paria block is considerably greater, causing the disparity in elevation of corresponding stratigraphic horizons in the two plateaus to increase in that direction.

Gregory and Moore (1931, p. 122–123) have described and mapped the monocline in southern Utah. They cite evidence for a strike fault along the monocline south of Paria. Here the Navajo sandstone shows indications of considerable fracturing accompanied by slipping and shearing. Furthermore, they point out, the extreme narrowness of House Rock Valley 10 miles south of Paria indicates a strike fault which cuts out part of the soft Triassic formation. They did not show the fault on their structure-contour map, and it has not been indicated on Plate 7. This locality was not studied by the present writers.

The northern end of the Kaibab Plateau resembles a highly asymmetrical north-pitching anticline. The West Kaibab fault, as it approaches the Arizona-Utah boundary, appears to turn into a gentle monocline. This flexure trends slightly more easterly than the East Kaibab monocline, causing the two structures to converge. From the State line northward, the portion of Kaibab Plateau between the two folds is reduced to the point where it is merely the crest of a highly asymmetrical anticlinal arch. The eastern limb is much steeper than the western, and the strata descend to far lower levels.

Great changes occur in both geological and topographical aspects of the fold where the Paria River crosses it. The Kaibab limestone anticline plunges beneath Triassic and Jurassic formations. The cliff of Jurassic sandstones which has paralleled the monocline base for many miles along the east side of House Rock Valley turns westward at Paria, encircling the plunging limestone arch, then continues southwest toward Kanab.

Except for the Paria River and Kaibab Gulch, drainage of the House Rock section is largely resequent and subsequent. Numerous small, subparallel, resequent streams have carved ravines and small canyons in the flank of the fold. These are tributary to House Rock Wash and Coyote Wash, axial subsequent streams which flow south and north respectively in House Rock Valley from a divide 8 miles north of House Rock. On the east wall of the valley short obsequents drain the Vermilion Cliffs. Upon the Paria Plateau drainage is northward by subparallel resequent streams flowing to the Paria Canyon across Jurassic Navajo sandstone.

Although the minor drainage falls into clearly recognizable classes, origin of the two principal drainage lines, the Paria River and Kaibab Gulch, is highly problematical. Only the latter stream will be treated here because it crosses that part of the monocline discussed. The stream originates in the southeastern slopes of the Paunsaugunt Plateau and flows southeastward across the White and Vermilion

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cliffs to the Kaibab arch. Four miles north of the State boundary the stream enters a deep, narrow canyon cut through the Kaibab arch and East Kaibab monocline at right angles to their strike. Near the center of the Kaibab Plateau the canyon is about 1500 feet deep. The gulch crosses House Rock Valley with no appreciable deflection of its course, enters a canyon in Navajo sandstone of the Paria Plateau, and 8 miles east of House Rock Valley joins the Paria River.

The southeastward course of the headward portion of the stream, down the terrace steps of the plateaus, is readily explained as an obsequent. Like Johnson and Kanab creeks farther west, it seems to have grown headward as the cliffs retreated northward. Davis (1901, p. 153–154) discussed the obsequent origin of these streams. The east-trending course of Kaibab Gulch, cutting across the arch and the monocline, is not so easily explained. Superposition of the stream from a cover of Eocene strata mantling the beveled Mesozoic strata is a possibility, but the northward dip of Eocene strata in the Paunsaugunt and Table Cliffs plateaus suggests initial consequent drainage northward rather than eastward.

Development of Kaibab Gulch can perhaps be explained by headward stream growth in Mesozoic rocks. From the crest of the monocline, eastward across House Rock Valley, the stream flows with the steep monoclinal dip and may have originated as a resequent stream, perhaps on the upper surface of the Navajo sandstone which now forms the Paria Plateau to the east. Because the Kaibab arch is highly asymmetrical with a very gently dipping west limb, the east-flowing resequent stream, which was then deeply intrenched in the monocline and cutting at a lower level than the land surface on the west limb of the arch, could have extended itself as an obsequent across the Kaibab arch, thence northwestward against the gentle dip of the strata. The deep subsequent lowlands on either side of the Kaibab arch developed later, after the thick series of Triassic shales had been exposed. It is not likely that under the present topographic conditions a stream could grow headward across even a low arch of hard rock such as the northern end of the Kaibab Plateau; instead, the weak rock belt swinging around the northern end of the plunging structure would be utilized. When the region was underlain by 1500 to 2000 feet of massive sandstones, however, and such weak-rock belts were not developed, conditions may have favored the drainage development suggested above. The course of the Paria River across the East Kaibab monocline may have developed in a similar manner.

PARIA DIVISION

GENERAL DESCRIPTION

The Paria Division is the northern end of the structure, extending from the town of Paria to the Paunsaugunt Plateau at Bryce Canyon National Park (Figs. 1, 15). This part of the monocline, as well as the adjacent area, has been described in detail by Gregory and Moore (1931). The present discussion is largely a restatement of observations recorded in their paper, with some additional material from the present writers' field studies. The areal geology and structure map (Fig. 15) is a combined and simplified copy of geologic and structure maps prepared by Gregory and Moore (1931). For stratigraphic data and excellent photographs the reader is referred to their report.

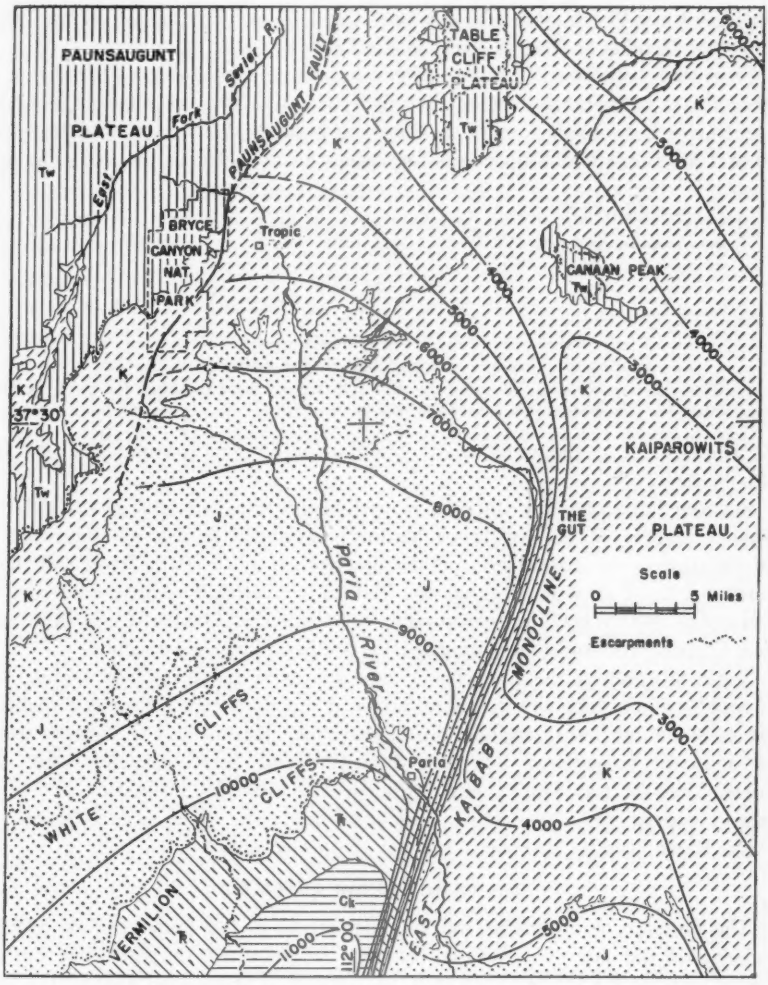

FIGURE 15.—*Geologic map of Paria division*

Structure contours on the top of the Dakota ss. Interval 1000 feet. Ck—Kaibab fm. Tr—Triassic. J—Jurassic. K—Cretaceous. Tw—Eocene Wasatch fm. After Gregory and Moore (1931).

At Paria, where the Paria River crosses the flexure, the monocline is a single, simple fold downthrown about 5000 feet on the east. It trends N. 25° E.; maximum dip is 45°. It is structurally continuous with the House Rock section, but, because of the northward regional dip, the fold is expressed in Jurassic and Cretaceous rocks rather than in Permian and Triassic rocks. Just north of the Paria River the monocline is expressed in hogback ridges separated by narrow subsequent valleys. From west to east are ridges of Navajo sandstone, Dakota sandstone, and Straight Cliffs sandstone. Subsequent valleys are developed on Chinle shale, San Rafael sandstones

and shales, and Tropic shale. Cottonwood Creek follows a valley in the San Rafael group for 15 miles as it flows south to the Paria River.

From Paria northward 15 miles to The Gut, the monocline shows little change in structure or appearance, but the stratigraphic units are progressively younger. The Chinle and Navajo formations disappear beneath the surface, while the Cretaceous Wahweap and Kaiparowits formations are included in the fold (Fig. 16).

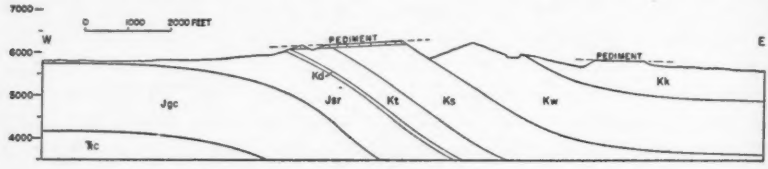

FIGURE 16.—*Structure section of East Kaibab monocline at The Gut*

Trc—Chinle fm. Jgc—Glen Canyon group. Jsr—San Rafael group. Kd—Dakota fm. Kt—Tropic fm. Ks—Straight Cliffs fm. Kw—Wahweap fm. Kk—Kaiparowits fm.

At The Gut the maximum dip is 35°, but the displacement is still nearly 5000 feet (Pl. 9, fig. 2).

Between Paria and The Gut, the monocline separates two geologically different plateau blocks. West is a broad platform developed on Navajo sandstone at an elevation of about 6000 feet, and deeply trenched by the Paria River and Hackberry Creek. East is the Kaiparowits Plateau, consisting here of broad structural terraces at 5000 to 6000 feet elevation on the Cretaceous Straight Cliffs, Wahweap, and Kaiparowits sandstones, trenched by southeast-flowing streams.

North of The Gut the monocline swings northwest and turns into a broad fold of rapidly diminishing throw. Near Tropic it can be detected only as a slight increase in the northward regional dip of the upper Jurassic and Cretaceous strata, but even this indication is lost along the east side of the Paunsaugunt fault. Between The Gut and Tropic beds dip north with sufficient inclination to produce ragged cuestas with intervening subsequent lowlands. The headwaters of the Paria River bear an obsequent relation to this structure.

AGE OF EAST KAIBAB MONOCLINE

The presence of Eocene as well as older strata in the Paria division provides evidence of the date of monoclinal flexing. Unfortunately the monocline dies out before reaching the Paunsaugunt fault, so it cannot be ascertained whether the fault, which cuts the Eocene Wasatch formation, also cuts across the East Kaibab monocline.

Canaan Peak, an erosional outlier of the Table Cliff Plateau, is capped with Eocene Wasatch strata resting upon the upper Cretaceous beds (Fig. 15). Although Canaan Peak does not lie on the axis of the fold, it is within the belt where the strata still possess a distinct dip. Gregory and Moore state (1931, p. 131) that here the eroded surface of the East Kaibab monocline is unconformably overlain by Eocene sediments, and also that (1931, p. 122–123) this unconformable relation suggests that the monoclinal flexing took place at the end of Upper Cretaceous time, before Eocene deposition. The present writers saw the unconformity described by Gregory and Moore on the southwest side of Canaan Peak. North-dipping Cretaceous beds

have here been truncated and are unconformably overlain by Eocene beds. On the basis of this evidence it is concluded that the East Kaibab monocline at least near this end was produced during the Laramide Revolution. Stratigraphic evidence with which to definitely date the remainder of the monocline is lacking, but the same date of deformation may reasonably be assigned to it because the monocline is a continuous structure, and nowhere was evidence of more than a single period of flexing found. This conclusion substantiates the belief of Walcott (1890), Davis (1901), and others that the East Kaibab monocline is an ancient structure, probably of early Tertiary age, and represents a period of deformation distinct from, and much earlier than, the period of great faulting of the High Plateaus of Utah and the western Grand Canyon district.

SUMMARY

The East Kaibab monocline, extending from the San Francisco Peaks in Arizona to Bryce Canyon in Utah, is a compound flexure downthrown on the east side. The southernmost or Cameron-Wupatki division terminates at its southern end in two branches paralleled by faults. Many small grabens and minor single faults are also present in this area. Quaternary lavas from the San Francisco Peaks volcanic field have usually flowed across the fault lines, but occasional minor faults break the lavas. Near Coconino Point, the two branches of the monocline join to form a single flexure with a throw of 1500 to 2000 feet and with dips ranging up to vertical. The various elements of the monocline, together with grabens, single faults, and rows of cinder cones, fall into three distinct trends—northwest, northeast, and north. Transitions in strike are accomplished by sharp bends. The Cameron-Wupatki region is underlain by the maturely dissected stripped surface of the Kaibab limestone, partly covered by volcanics. Pediment remnants cut in weak Triassic shales are present along the monocline base.

In the Grand Canyon area, the monocline again branches. The main line of displacement runs northward through the eastern end of the Grand Canyon where it is paralleled by the Butte fault. A minor line of flexing, which stems from the main one south of the Canyon, runs northwestward into the Canyon and dies out north of the Colorado River. The river follows the strike of major structures closely and appears to have originated either as a subsequent stream in a weak-rock belt or, if originally superposed or antecedent, has since become closely adjusted to structure.

North of the Grand Canyon the monocline forms the eastern margin of the Kaibab Plateau, dropping the strata down to the level of the Paria and Marble Platform blocks. Part of this division is a double flexure with a tread between, but elsewhere it is a single fold. The throw increases from 2500 feet near the Grand Canyon to 4000 feet in southern Utah. The northern part is paralleled on the east by House Rock Valley and the Vermilion Cliffs. Extensive pediments lie along the base of the monocline near the southern end of House Rock Valley.

The northernmost or Paria division of the monocline extends from Paria to Bryce Canyon, Utah. Here the monocline is a single fold of about 5000 feet throw in Jurassic and Cretaceous strata. Near Bryce Canyon the monocline bends westward and dies out rapidly. The Eocene Wasatch formation unconformably overlies the tilted upper Cretaceous strata, indicating Laramide deformation for the monocline.

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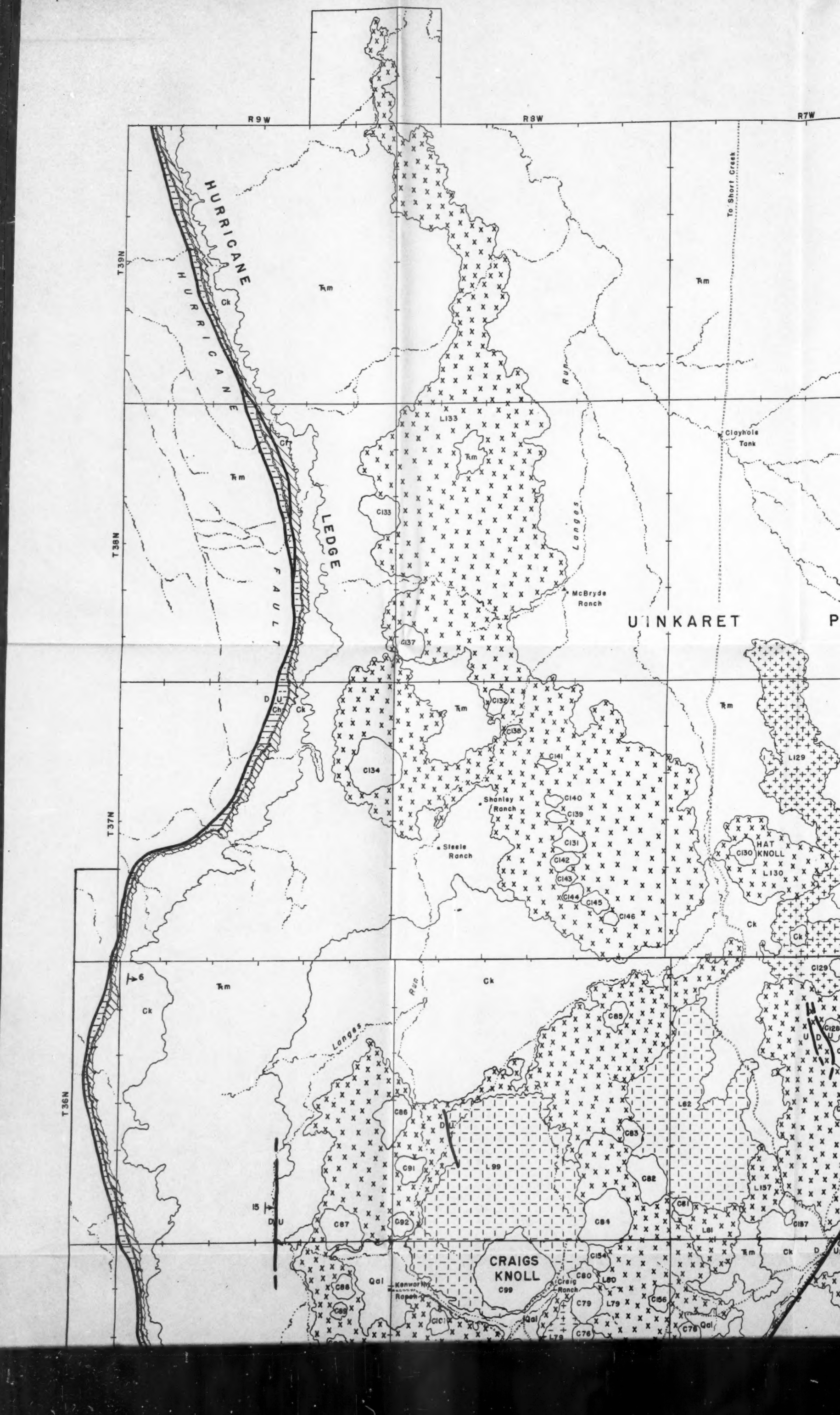
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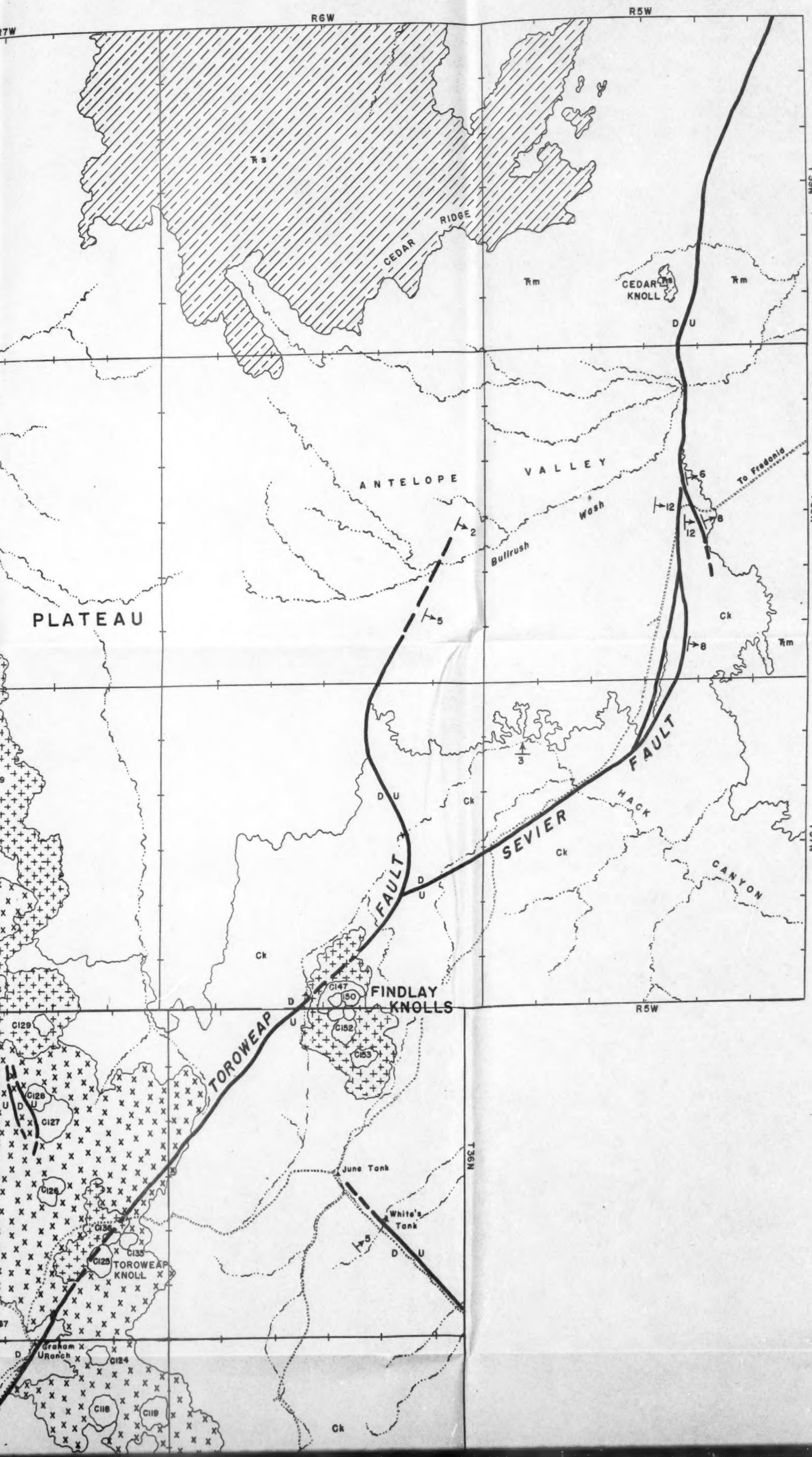
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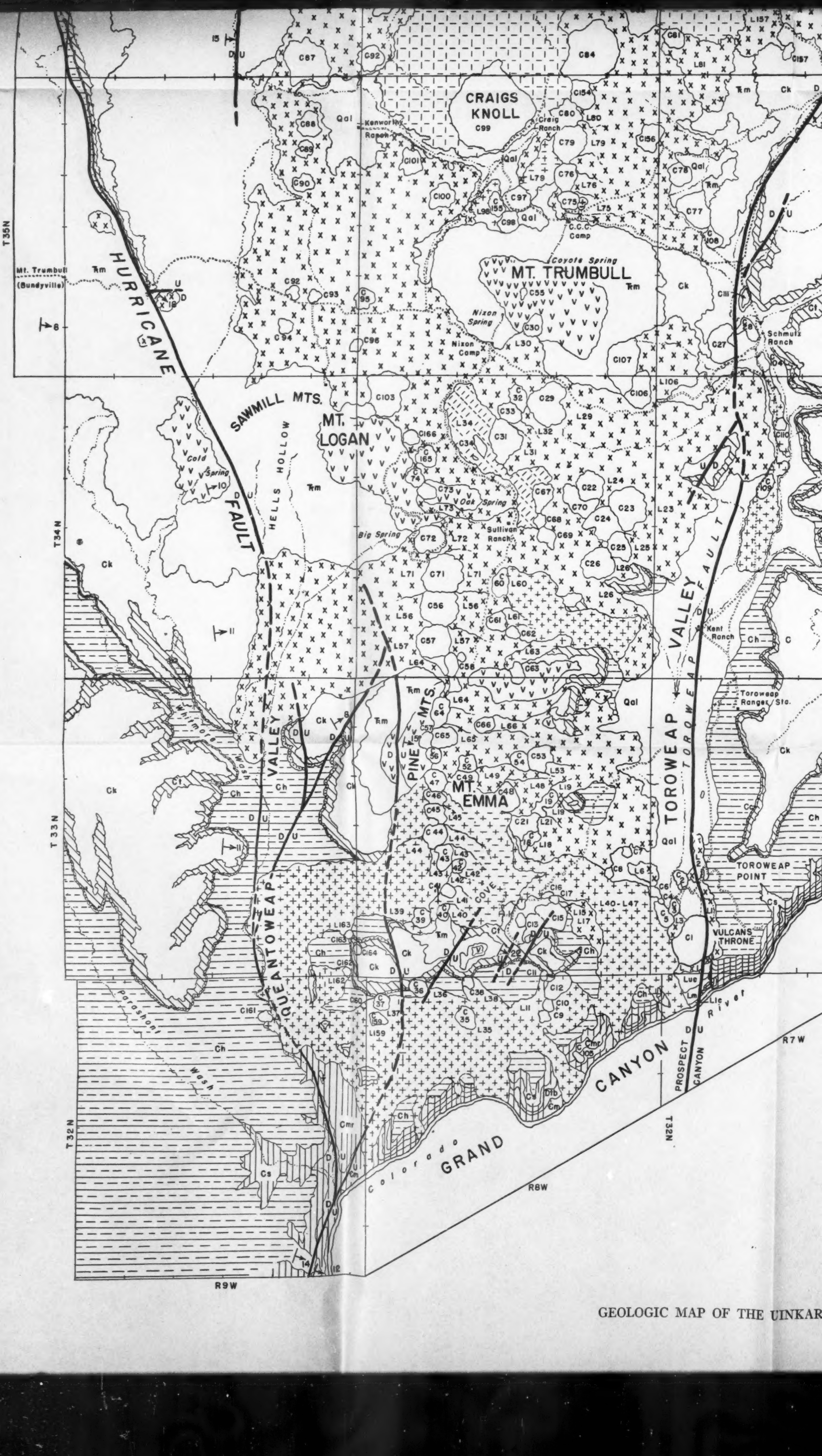

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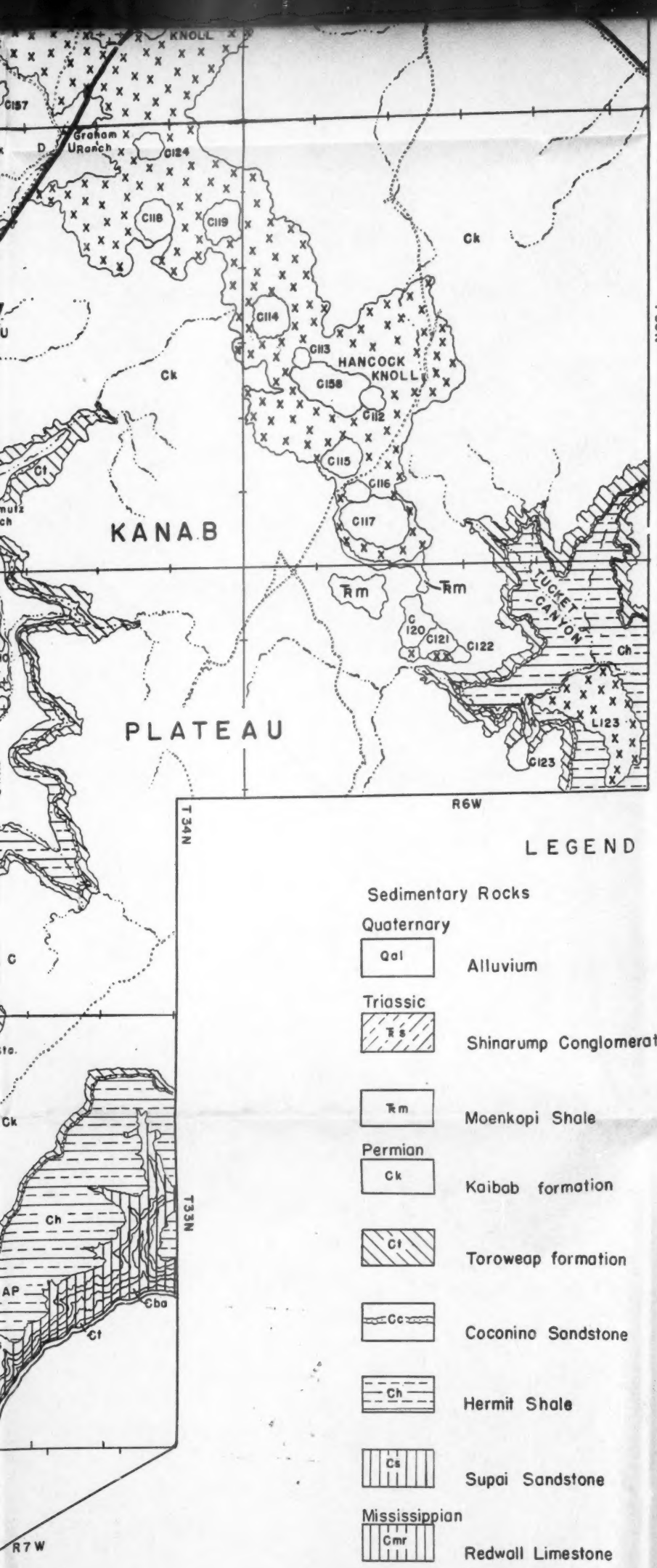

LEGEND

Sedimentary Rocks

Quaternary

- Qal — Alluvium

Triassic

- Ts — Shinarump Conglomerate
- Tkm — Moenkopi Shale

Permian

- Ck — Kaibab formation
- Ct — Toroweap formation
- Cc — Coconino Sandstone
- Ch — Hermit Shale
- Cs — Supai Sandstone

Mississippian

- Cmr — Redwall Limestone

Devonian

- Dtb — Temple Butte formation

Cambrian

- Cm — Muav Limestone
- Cbc — Bright Angel formation
- Ct — Tapeats Sandstone

Lavas of the Inner Gorge of the Grand Canyon

- Luc — Upper Canyon Group
- Lmc — Middle Canyon Group
- Llc — Lower Canyon Group

Igneous Rocks of the Uinkaret and Kanab Plateaus

- C 147 — Cinder Cones
- Stage IV Lava
- Stage III-c Lava
- Stage III-b Lava
- Stage III-a Lava
- Stage II Lava
- Stage I Lava
- C133 — Cone numbers
- L133 — Flow numbers
- D / U — Faults
- 22 — Strike and dip
- Intermittent stream
- Tank
- Pipe line
- Spring

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Scale

UINKARET PLATEAU

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GEOLOGY OF THE UINKARET PLATEAU, NORTHERN ARIZONA

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ABSTRACT

The Uinkaret Plateau, a major block of the Colorado Plateau province in northern Arizona, is bounded on the west by the Hurricane Ledge, developed along the Hurricane fault, and on the east by scarps developed along the Toroweap and Sevier faults. Over the southern part there are extensive outpourings of lava which fall into four age groups.

The oldest eruptions preceded earliest movement along the Hurricane fault and antedated cutting of the present Grand Cayon. These lavas poured out on a relatively even surface believed to have been a large pediment or series of pediments extending at least 16 miles north from the Colorado River. Extensive dissection of the oldest lavas preceded the next eruptions, which took place at considerably lower elevations. The Inner Gorge of Grand Canyon was cut to approximately its present depth and later filled to a height of at least 600 feet and perhaps 1,200 by lava floods from Toroweap Valley. The lava dams were entirely removed before later flows, pouring into the Inner Gorge from the rim of the Esplanade, again dammed the river, which has removed all but small portions of these dams. Repeated movements along the Toroweap fault have occurred in post-Iowan-Wisconsin time. In the very recent past, renewed volcanic activity formed a single small cone and lava flow.

INTRODUCTION AND ACKNOWLEDGMENTS

The Uinkaret Plateau is one of several great blocks constituting the western Colorado Plateau province in northern Arizona and southern Utah. In this paper the writer (1) describes geological and topographical features of the Uinkaret Plateau and immediately adjacent areas; and (2) interprets the geomorphic history of this region.

The author acknowledges his indebtedness to the late Professor Douglas Johnson, who directed the project, critically read the manuscript, and offered valuable suggestions; to Dr. Arthur N. Strahler and other members of the Columbia Seminar in Geomorphology for many useful criticisms; to Ranger William Bowen of Grand Canyon National Monument, who was a source of invaluable assistance and information in the field; to Professor Edwin D. McKee, Assistant Director of the Museum of Northern Arizona, and other members of the Museum staff who were generous with their time and the facilities of the organization; and to Mr. Lawrence Curtis for field assistance during the summer of 1941.

PREVIOUS WORK

Peaks of the Uinkaret volcanic group were sighted by members of the Ives Survey, 1857–58, but their explorations were confined to the country south of the Colorado River. Ives (1861, p. 104, 110, and map no. 2) referred to the peaks as the "north-side mountains." The first recorded survey of the Uinkaret field was made by Powell (1875) in 1869–72. He applied the name Uinkaret to the plateau block as a whole, and the names Mt. Trumbull, Mt. Logan, and Mt. Emma to the peaks. He discovered the Toroweap fault and estimated its displacement as 800 feet (1875, p. 94). Dutton (1882) studied the area in more detail. Reconnaissance surveys by Davis (1901; 1903) and Johnson (1909) brought further contributions.

More recently, detailed studies have been made by Moore (1925), McKee (1938), and McKee and Schenk (1942), whose examinations were limited principally to local stratigraphy or to detailed studies of very small areas.

FIELD WORK

The writer spent most of the field seasons of 1941 and 1942 in the area. The recent topographic map of the Grand Canyon National Monument made by the U. S. Geological Survey, aerial photographs from the Department of Agriculture, Forest Service maps, and General Land Office maps were compiled into the base map. Parts of the area for which no maps existed were mapped by plane table on a scale of 1:24,000.

The survey resulted in a detailed geologic map of the region (Pl. 1), division of the volcanic activity into four stages which can be partially correlated with those of the San Francisco Peaks volcanic field, reinterpretation of structural details, dating of the major displacements, and reinterpretation of the geomorphic history. No evidence of the peneplane previously thought to have been produced by the "Great Denudation" was found; on the contrary, evidence indicates that the history has been one of normal single-cycle arid-climate erosion, with extensive pedimentation.

GENERAL DESCRIPTION ON UINKARET PLATEAU

The Uinkaret Plateau extends northward 50 miles from the Colorado River to the Vermilion Cliffs and Virgin River in Utah (Fig. 1). The western boundary is the Hurricane Ledge, a scarp mostly 1,500 to 1,800 feet high, which marks the Hurricane fault. The eastern boundary, less sharply defined, consists of Toroweap Valley, developed along the Toroweap fault, and the low escarpment which in places marks the Sevier fault. The plateau widens from about 8 miles at the Colorado River to over 35 miles in the north. On the central and southern parts numerous lava flows issued from more than 160 craters distributed over the plateau (Pl. 2, fig. 1).

Elevations along the western margin of the Uinkaret Plateau range from 6,000 to 7,000 feet; along the eastern margin, from 5,000 to 5,200 feet. Rising mesalike above the level of the plateau are the tabular masses of Mt. Trumbull, 8,028 feet, Mt. Logan, 7,700 feet, and the Pine Mountains, 7,650 feet (Pl. 1). The floor of Queantoweap Valley, developed along the Hurricane fault, lies at elevations of 3,000 to 5,000 feet and is partially underlain by lava flows from cones north and south of the Pine Mountains 5 miles east of the valley. The floor of Toroweap Valley is about 1,000

feet higher than Queantoweap Valley and is almost entirely underlain by lavas which flowed into the valley in five great cascades from cones on the plateau.

The region is semiarid, typical of the lower portions of the Colorado Plateau province, but marked local variations occur because of the wide range in elevations. Thus, within the Inner Gorge of Grand Canyon, the climate is that of the Lower Sonoran Life Zone, whereas on Mt. Trumbull, Mt. Logan, and Mt. Emma conditions are characteristic of the temperate Transition Life Zone.

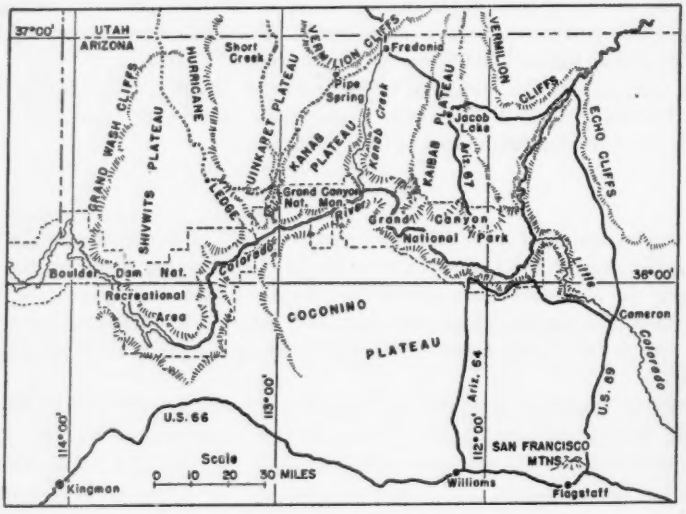

FIGURE 1.—*Index Map of Grand Canyon district*

STRATIGRAPHY

At the bottom of the Grand Canyon east of the Toroweap fault the Colorado River now flows in Cambrian Tapeats sandstone with approximately 70 feet exposed (Pl. 1). Overlying the Tapeats is the Middle Cambrian Bright Angel formation, consisting of 600 feet of shale and interbedded limestone. The Bright Angel is followed by the Cambrian Muav limestone, 470 feet thick, forming the lower cliff of the Inner Gorge (Pl. 3, fig. 3). This is overlain by the Devonian Temple Butte formation, 335 feet thick, and massive Mississippian Redwall limestone 600 feet thick, which forms a sheer cliff (Pl. 3, fig. 3). The succeeding Permian Supai formation, 616 feet thick, consists of cross-bedded sandstones with some limestone near the base.

These formations are exposed only in the Inner Gorge of Grand Canyon and in tributary canyons. Above the Supai the Hermit formation, 1,053 feet thick, contains in the lower 200 feet thick cross-laminated beds of sandstone on which the Esplanade, the floor of the upper, outer canyon, is developed. The upper 800 feet of the formation is composed of weak shales and sandstones forming slopes below the outer walls of the upper canyon (McKee and Schenk, 1942). Next is the massive cross-bedded Permian Coconino sandstone, 100 feet thick in the eastern part of the area, thinning westward to 50 feet along Queantoweap Valley. The overlying Per-

FIGURE 1. NORTHERN PORTION OF UINKARET VOLCANIC FIELD
Viewed from Graham Ranch. North slope of Mt. Trumbull at far left. At right, cone 82, a large composite Stage II cone.

FIGURE 2. CONE 102
Volcanic neck (left of center) and dike (left). Erosional remnant of cone at upper right. Note absence of disruption of strata adjacent to neck.

CONES OF THE UINKARET PLATEAU

FIGURE 1. VIEW SOUTHWESTWARD DOWN INNER GORGE OF GRAND CANYON
Lava cascades in center have flowed from Toroweap Valley mouth at upper right. Similar lava cascades in distance. (Photograph by A. N. Strahler.)

FIGURE 2. MIDDLE CANYON LAVA
Seen adhering to wall of inner gorge of Grand Canyon, half a mile east of Toroweap fault. Top of mass is approximately 1200 feet above Colorado River.

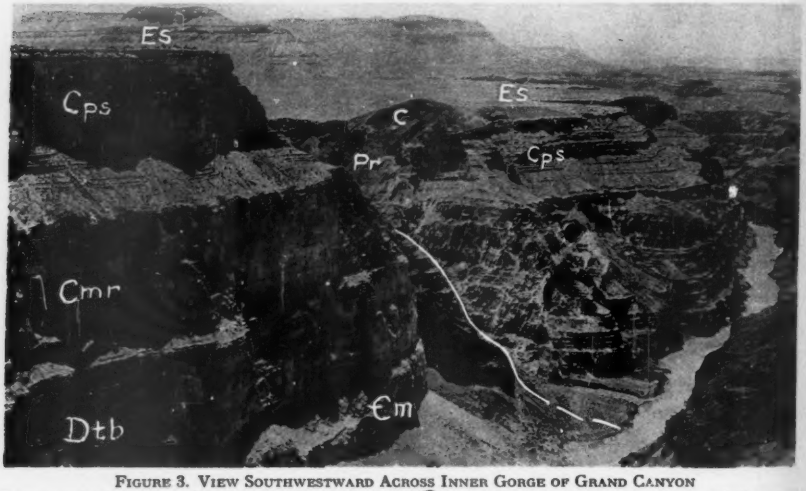

FIGURE 3. VIEW SOUTHWESTWARD ACROSS INNER GORGE OF GRAND CANYON
From a point just east of Vulcans Throne. White line shows position of Toroweap fault. Upthrown block nearest the observer. Cm-Muav limestone, Dtb-Temple Butte limestone, Cmr-Redwall limestone, Cps-Supai sandstone, Es-Esplanade, Pr-Prospect Canyon, C-cone. (Photograph by A. N. Strahler.)

GRAND CANYON AT FOOT OF TOROWEAP VALLEY

mian Toroweap formation consists of 425 feet of interbedded limestone, shale, and sandstone, with many intercalations of gypsum. The Permian Kaibab formation, 570 feet of interbedded limestone, sandstone, and shale, forms the uppermost part of the outer Grand Canyon walls (McKee, 1938). The stripped upper surface of this resistant limestone forms most of the southern Uinkaret and Kanab plateaus not covered by lava.

Small patches of Triassic Moenkopi shale, generally under a protective covering of lava, overlie the Kaibab. The entire 1,600 feet of the formation is preserved beneath the lava cap of Mt. Trumbull. The Triassic Shinarump conglomerate is the youngest sedimentary formation outcropping in the area studied, where it ranges from 50 to 100 feet thick.

IGNEOUS ROCKS

COMPOSITION

Perhaps the most remarkable feature of the lavas of the Uinkaret volcanic field is their similarity in composition and appearance. Despite great ranges in age, the lavas are all essentially fine-grained olivine basalts. The groundmass is very fine-textured, consisting of 50 per cent or more plagioclase, the balance being augite, magnetite, and some olivine. Occasional large insets of olivine and euhedral augite are found. In the upper portions of flows and near cones are masses of olivine up to 6 inches in diameter. There are subhedral hornblende crystals up to 1½ inches long, characterized by very low pleochroism and absence of any marginal alteration. Greater variation of composition and texture were found both vertically and horizontally within one flow than between widely separated flows. Thus there is no possibility of correlation of partially buried flows or of those within the canyon with others in the field on the basis of composition alone.

The lavas are dark gray when fresh and weather to a lighter gray, sometimes with a brownish cast. Few flows are more than 25 to 30 feet thick, and many show excellent columnar jointing. The upper surfaces of more recent flows, and of older flows where protected by overlying lavas, are very scoriaceous. Where flows occur in series, they are usually separated by beds of volcanic ash 6 inches to 6 feet thick.

Individual flows, when they can be distinguished from the larger coalescent sheets, are generally of small areal extent, though rare examples, such as L 133 (Pl. 1), extend over 12 miles from the cones. The grouping of contemporaneous cones has resulted in large lava sheets composed of many flows coalescing laterally, sometimes consisting of 10 or 15 sheets separated by thin layers of cinder and ash. Such lava floods formed the great cascades which flowed into Queantoweap and Toroweap Valleys from the plateau between.

The cones are largely volcanic ash and cinders, indurated in the interior but loose on the surface. Except the most recent cone, all are characterized by the red of the scoriae. Bombs from 2 inches to more than 3 feet long are common. The cones range from 50 to over 800 feet in height, and 27°–30° in slope. Craigs Knoll is the largest. All the cones have been breached by the most recent lava eruptions and are therefore asymmetrical. In a few deep erosion has exposed their internal structure. Cone 162 has been sufficiently eroded to expose the central neck (Pl. 2, fig. 2).

RELATIONS OF CONES TO STRUCTURE

Most volcanic eruptions are thought to occur along major lines of weakness. One would thus expect to find cones associated with the three major fault lines of this area and with the numerous minor structures of the plateau and adjoining region. The map (Pl. 1) shows that this is not so. Dutton (1882, p. 105) remarked on the tendency of cones to appear between major displacements rather than along them. Detailed mapping confirms his views. Eruptions along fault lines appear to be coincidental. Thus Volcans Throne, Toroweap Knoll, Findlay Knolls, and cones 27, 28, 104, 111, 109, 110, and 161 are the only ones on major fault lines. In contrast, 23 cones east of the Pine Mountains and Mt. Logan occur along a northward-trending line; 22 cones, from cone 123 to Hat Knoll 12 miles north of Mt. Trumbull, occur in a northwestward-trending line; 11 cones west of Hat Knoll have a similar trend; and 6 extend northward from Mt. Trumbull. There are other less striking examples of alignment. Detailed examination of exposed bedrock near these aligned groups shows that, though the lines of cones parallel fault trends the cones do not lie on observed fault lines. Thus, excepting those mentioned, cones tend to occur in distinct lines, parallel to, but not associated with, observable surface fractures. This parallelism may have resulted from stresses in existence at the time of deformation. Professor Walter H. Bucher (personal communication) suggested that, since the fault planes are somewhat inclined and consequently tight, they would offer the least favorable path for eruptions, which would be more likely to use relatively open joints near the fault line. Other examples of eruptions occurring near, but not on, major fault lines are reported from the Rhine Graben and the Rift valleys of Africa. It is, of course, possible that zones of weakness at depth, having no surface expression, are responsible for the alignment.

DETERMINATION OF AGE RELATIONS OF LAVAS

Powell separated the Uinkaret basalts into two principal groups (1875, p. 131–132), the ancient basalts which underlie the Trumbull, Logan, and Emma platforms, and the much younger group which forms the bulk of the field. Dutton (1882, p. 110–111) further divided later flows into middle-aged and young. This broad classification is based on topographic relationships; earlier flows are limited to the high mesas, while later flows generally occur at much lower elevations. This indicates a considerable time interval separated the two epochs. The later flows are divided according to surface appearance, the middle-aged having suffered much more extensive erosion and weathering than the younger.

If the rate of weathering of flows and cones is proved approximately uniform throughout the field, it follows that time has been the essential factor in determining amount of weathering and erosion of different flows. A classification based on extent of weathering and erosion could then be employed, and correlations made between flows widely separated in the field. Colton proposed (1937) such a classification for the San Francisco Peaks volcanic field to the southeast.

Factors influencing the rate of such weathering and erosion include composition, texture, size of fragments, precipitation, evaporation, vegetation, temperature changes, frost action, and surface slope.

Composition, texture, and fragment size are essentially uniform throughout the

area; hence no systematic variations in the rate of weathering or erosion attributable to these factors can be recognized. In the Uinkaret field, no data on climatic conditions are available; hence only generalizations of the effects of precipitation, evaporation, insolation, vegetation, and frost action can be made. In the vertical range from 1,650 feet at the bottom of the Inner Gorge to 8,028 feet at the top of Mt. Trumbull there are four climatic zones: the Lower Sonoran Life Zone, the Upper Sonoran Life Zone, the Transition Life Zone, and the Canadian Life Zone. Precipitation ranges from less than 8 inches to more than 30 inches annually. Frosts are rare at low elevations but frequent at high elevations. Evaporation and insolation also vary with elevation. The vegetation changes both in type and density.

Effects of these factors are pronounced over long periods of time, but for the relatively short time the flows have been exposed variations are slight. Thus a single flow may be traced from the lower Sonoran zone to the Transition zone, with attendant changes in climatic conditions and vegetation cover, without exhibiting marked changes in weathering. Another lava sheet may be traced from the upper Sonoran zone through the Transition zone with very little change in appearance. Thicker vegetation at higher elevations only tends to mask the appearance of flows from a distance.

Vulcans Throne in the upper Sonoran zone is associated with a flow correlated with another which may be traced to a cone in the upper Transition zone. Despite marked differences in all climatic conditions, the weathered surfaces of these two cones are remarkably similar. Duration of exposure thus seems to be more effective than climatic changes in determining extent of weathering.

Absence of appreciable variations resulting from steepness of slope alone indicates only slight influence of that factor. Single flows may be traced down slopes of more than 20° to almost horizontal surfaces without noticeable change. Slopes of cones are of the same order of magnitude regardless of age or size; hence the effect of this factor is substantially the same on all cones.

While these factors may have pronounced effects over a long time, the length of time the lava has been exposed is so short that they caused no notable differences in weathering. The extent of weathering and erosion appears to depend chiefly on the length of time individual cones and flows have been exposed. A satisfactory classification by relative ages of cones and flows may thus be based on their surface appearance. This correlation, when coupled with stratigraphic correlation at a type locality, is believed to provide a valid classification of the cones and flows of the entire field.

CLASSIFICATION OF CONES AND FLOWS

Nowhere in the Uinkaret field are flows of all ages in contact with each other, but flows L 15 and L 16, L 40 to L 47, and the old lavas of the Emma Platform, exposed in southern Toroweap Valley and tracable to parent cones on the plateau, provide a type section (Pl. 1). On the basis of their relative stratigraphic positions and extent of weathering and erosion, the lavas fall naturally into four major groups or stages.

The oldest or Stage I flows are so extensively weathered and eroded that all initial surface features are removed and original margins destroyed. They occur only as cap rocks of the isolated platforms of Mt. Trumbull, Mt. Logan, and Mt. Emma.

No cones are associated with them, though the north face of Mt. Trumbull exhibits in cross section what appears to be a volcanic neck (Fig. 2).

Stage II flows, restricted to the northern part of the field, have also been extensively weathered and eroded, and their initial surface irregularities removed. Original flow fronts are not preserved, although modification of margins does not appear to have been extensive except where underlain by weak rock. There are deeply

FIGURE 2.—*Sketch of north face of Mt. Trumbull*
Showing structure which is interpreted as a neck. (Drawn from a field sketch.)

gullied cones, for example, Craigs Knoll, 3 miles north of Mt. Trumbull, and cones 82 and 103 (Pl. 1). Stage II flows of cone 103 are almost entirely buried beneath more recent flows. These three cones differ from younger ones in that they are large and composite. Possibly the more extensive erosion of Stage II cones, as contrasted with later ones, resulted from greater initial irregularities of shape, which in turn concentrated run-off and led to more rapid dissection than in more simple cinder cones. Possibly, then, the cones of Stage II appear older than they actually are.

Most lavas of the Uinkaret field belong to the third stage of activity, characterized by spatter cones on the flows, some marginal modification by streams, and moderate weathering of the surface. This may further be subdivided into Stages III-a, III-b, and III-c. Though the flows of the substages are characteristically different and readily distinguished, it is not possible to separate cones into three stages. Usually two substages of activity seem to have taken place at the same vents, and in a few cases, all 3 substages are represented.

The surfaces of Stage III-a flows have been broken down by weathering, but spatter cones are occasionally visible. The margins are frequently outlined by stream channels consequent between the lava and adjacent slopes, which sometimes modify the boundaries of the sheets. Where III-a flows lie on weak rocks, the margins have sometimes retreated a considerable distance. Where flows reach the floor of Toroweap Valley their margins are buried under alluvium, so that there is rarely a distinct topographic break between flow and valley fill.

The margins of Stage III-b flows have been slightly weathered and eroded. The surfaces are somewhat altered and there is sufficient soil to support scanty vegetation. Spatter cones are common and well preserved. Margins of these flows in Toroweap Valley are only partially buried by alluvium, resulting in a distinct topographic break between flow and valley floor.

Freshness of surface and outline marks Stage III-c flows, though there is sufficient soil to support sparse vegetation. These flows are of limited areal extent and appear to be products of late activity from vents which produced III-a and III-b flows.

The single Stage IV flow, L 34, 3 miles south of Mt. Trumbull, is remarkable for its fresh surface and appearance of recency (Pl. 1). Even the finest material of the cone is only slightly altered. Dutton (1882, p. 111) says, "It looks as fresh as any *coulee*

of Vesuvius ejected twenty or thirty years ago." Despite the apparent freshness of the flow, no legends of recent eruptions are reported among the Indians of the district, thought by anthropologists to have reached the area 900 to 1000 years ago. This suggests that the outbreak is pre-Paiute Indian.

Flows of the type section are classified as follows: L 15 and L 16, Stage III-a; L 40 to L 47, Stage III-b; Cove flows, Stage III-c; Emma Platform, Stage I (Pl. 1). From this area, correlations may be carried northward to Stage II and Stage IV flows, completing the classification.

CORRELATION WITH INNER CANYON LAVAS

McKee and Schenk (1942, p. 251–252) recognized four groups of Inner Canyon lavas (Pl. 1; Pl. 3, fig. 2): (1) the Lower Canyon group, consisting of 727 feet of interbedded lava, ash, and river-deposited sediments; (2) the Middle Canyon 900 feet of interbedded lava and ash; (3) the Upper Canyon 900 feet of lava; and (4) the Valley Surface flows, which cascaded over the front of the Inner Canyon lavas (Pl. 3, fig. 1).

Absence of any physical connection of the Inner Canyon lavas with those of the Uinkaret Plateau renders correlation difficult. The composition of the Inner Canyon and Uinkaret lavas is similar, as is the general appearance of the flows. Possibly the Lower Canyon lavas were of local origin and have no connection with others of the region. Presence of the Lower Canyon group at the bottom of the Inner Gorge, and their horizontal attitude, suggests that they are considerably younger than Stage I flows. Other factors, discussed later, confirm this conclusion.

Lavas of the Middle Canyon group appear to have flowed for an undetermined distance down the floor or Toroweap Valley before entering the Inner Gorge (McKee and Schenk, 1942, p. 264). Absence of any Stage II cones in the immediate vicinity militates against their correlation with Stage II, but possibly Stage II vents in Toroweap Valley are buried by later flows. Surface exposure of the Middle Canyon group is insufficient to permit determination of age on the basis of weathering and erosion.

The younger flows of the Upper Canyon group can be traced to Stage III-a flows in Toroweap Valley. Some appear to have originated from cone 2, others from Vulcans Throne, though it is not certain that all lavas of this group are derived from these cones. Some may well have come from vents on the Uinkaret Plateau.

In summary, no correlation of the Lower Canyon group with stages on the plateau is possible. Evidence is not conclusive concerning correlation of the Middle Canyon group, but the author suggests that these flows originated from Stage III-a vents north and west on the Uinkaret Plateau. The Upper Canyon group is of Stage III-a age, as indicated by physical connection with III-a flows north of Vulcans Throne. The Valley Surface flows belong to Stage III-b.

LAVA DAMS AT TOROWEAP

Damming of the Colorado River to a height of about 550 feet by Lower Canyon lavas has been demonstrated by McKee and Schenk (1942, p. 264–266). Powell (1875, p. 95) states that the canyon "was doubtless filled to a height of 1,200 or 1,500 feet" in this section, but gives no supporting evidence. Similarly, Moore (1925, p. 156) states that flows "undoubtedly dammed the lower part of the canyon to a

height of some hundreds of feet." Presence of the large mass of Middle Canyon lava (Pl. 3, fig. 2), the top of which is 1,200 feet above the river, half a mile east of the main mass of Middle Canyon lava, may indicate the minimum upstream distance to which those flows extended. A similar smaller hanging remnant, 1,200 feet above river level, is south of the river 2 miles east of the Toroweap fault. This does not, however, provide conclusive evidence of damming by middle canyon flows to 1200 feet above river level.

CORRELATION WITH NEIGHBORING FIELDS

No detailed surveys have been made of the volcanic field of the Shivwits Plateau (Fig. 1) west of the Uinkaret. Colton (1937) examined and classified San Francisco Mountain basalts on a basis similar to that applied in the Uinkaret field (Fig. 1) Reconnaissance surveys by the writer indicate that, because of similarity in texture and composition, and of climatic conditions, the effects of weathering and erosion are approximately the same in the two fields. Colton divides activity in the San Francisco field into five stages. Because of similarity in appearance and topographic position, Stages I, II, and III seem to be approximately contemporaneous in the two fields. Stage IV of the Uinkaret field appears to be of the same age as Stage V in the San Francisco field. No equivalent of the San Francisco field Stage IV is found in the Uinkaret area.

TOROWEAP, SEVIER, AND HURRICANE FAULTS

TOROWEAP FAULT

The Toroweap fault, one of the major displacements of the Colorado Plateau province, separates the Kanab and Uinkaret divisions (Fig. 1). From a point 7 miles north of the Findlay Knolls, 15 miles northeast of Mt. Trumbull, it extends southward down Toroweap Valley, thence across Grand Canyon, and up Prospect Canyon for an undetermined distance (Pl. 1). It is a normal fault with the eastern side upthrown.

Powell (1875, p. 94) first noted the Toroweap fault and estimated its displacement as 800 feet. Dutton (1882, p. 93–94) made a more detailed examination, determined the throw as between 600 and 700 feet, and noted the slight eastward inclination of the fault plane. Because the fault cuts basalts, and because recession of cliffs from the fault line is slight (Pl. 3, fig. 3), Dutton concluded that the movement was very recent.

Davis (1901, p. 142–143; 1903, p. 18–20) and Johnson (1909, p. 194) found evidence that the faulting was older than Dutton had supposed and that Toroweap Valley and Prospect Canyon had developed along the line of weakness. Davis also discovered evidence that movement along the fault took place at two widely separated intervals. Moore (1925, p. 156) determined the displacement as 580 feet but made no division into pre-lava and post-lava faulting. McKee and Schenk showed (1942, p. 262) that pre-lava faulting amounted to 486 feet and was accompanied by considerable drag on the downthrown block. Post-lava faulting of 146 feet is marked by absence of any crushing or drag, a fact also noted by Dutton.

There is evidence that not all this post-lava movement took place at the same time. A small fault scarp, not over 20 feet high, cuts the eastern side of Vulcans Throne

(Pl. 1). It is not likely that a scarp developed in unconsolidated cinders which compose the cone would exist unaltered for any great length of time, but it is apparent that the cone was never faulted to the extent of 146 feet. In Prospect Canyon,

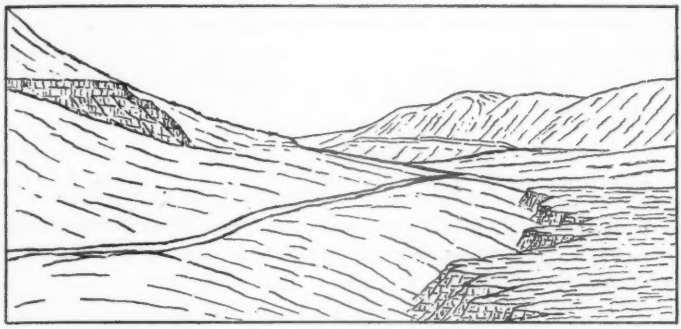

FIGURE 3.—*Toroweap fault in Toroweap Valley*

View southward along fault near Graham Ranch at head of Toroweap. Fault coincides approximately with road in center. Lava in upper left is hanging remnant on upthrown block. Corresponding downthrown lava on right.

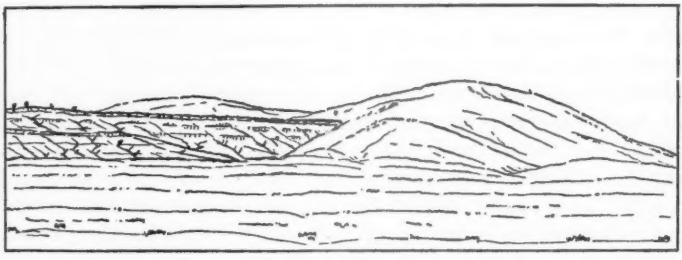

FIGURE 4.—*Toroweap fault at Findlay Knolls*

Toroweap fault-line scarp in Kaibab limestone, left from center. At right, Stage III-b flows from Findlay Knolls burying scarp.

south of Grand Canyon, Davis (1903, p. 20) described a small fault scarp 20 to 40 feet high and only slightly altered by erosion. Flows L 1 and L 2 north of Vulcans Throne are bounded on the west by a fault scarp 15 to 35 feet high. The foot of this scarp is buried in alluvium, making it impossible to determine the exact amount of displacement. Northward in Toroweap Valley Stage III-a flows L 29, L 30, and L 32 have been faulted 10 feet. This structure may be traced northward past cone 27, which is faulted on its eastern side. Five miles north of cone 27, flows issuing from the Hancock Knolls have been faulted 150 feet, leaving hanging remnants of flows on the eastern, upthrown block (Fig. 3). Displacement decreases northward for the next 4 miles, until the scarp disappears beneath unbroken III-b flows from Toroweap Knoll and cones 135 and 136. North of Toroweap Knoll the escarpment reappears, and again III-a lavas are cut. The scarp continues northeastward, passing below Stage III-b lavas from Findlay Knolls (Fig. 4). There is thus evidence of movement of 10 to 150 feet after extrusion of III-a flows, but before III-b eruptions.

A mile and a half north of Findlay Knolls, the Toroweap fault makes an angular junction with the Sevier fault (Pl. 1). At the junction, throw along the Toroweap fault is approximately 75 feet, that along the Sevier fault 25 to 35 feet. Beyond this junction, the Toroweap fault curves northwestward for 3 miles, then northeastward for approximately 2 miles, with decreasing displacement. Thus about 5 miles north of the junction it becomes a monocline with throw of slightly more than 50 feet, as described by Johnson (1909, p. 152), and continues northeastward for several miles until it disappears.

SEVIER FAULT

From its junction with the Toroweap fault, the Sevier fault swings eastward past the head of Hack Canyon, thence northward past Cedar Knoll to Pipe Springs and beyond into Utah, where it becomes one of the most prominent faults cutting the High Plateaus.

Davis (1901, p. 143) stated that the Toroweap and Sevier faults were continuous, not separate as Dutton had supposed, and applied to this displacement the name Toroweap-Sevier fault. Johnson (1909, p. 152) traced the Toroweap fault southward from its origin in a low monocline north of the Toroweap Valley and concluded that the Toroweap and Sevier faults were separate displacements.

Detailed surveys by the writer show that the Sevier fault curves sharply westward to join the Toroweap fault north of the Findlay Knolls. Had these displacements developed at different times, expectable consequences would be: (1) the Sevier fault might continue beyond the Toroweap fault; or (2) some flexing of the beds beyond the Toroweap fault would mark the westward continuation of the Sevier displacement; (3) some flexing of the upthrown beds in the immediate vicinity of the fault might have occurred. There is no indication that the Sevier line of displacement continued beyond the Toroweap fault, either as a fault or fold, nor are the beds of the upthrown block flexed at the junction (Fig. 4). The writer therefore concludes that movement along this portion of the two faults was contemporaneous. Thus, although the Toroweap and Sevier faults are separate lines of displacement, as Johnson stated, they do join, as postulated by Davis.

At the junction of the two faults, throw along the Sevier fault amounts of 25 to to 35 feet, with the eastern side upthrown. This increases to slightly more than 200 feet at Hack Canyon, beyond which two branches of the fault appear, each having a displacement of more than 200 feet, with the intervening block tilted 15° E. Six miles north of Hack Canyon, where the road to Fredonia crosses the fault line, another branch appears, which extends southeastward for 3 miles. Along the margin of the upthrown block, the Kaibab beds dip 8°–12° E. At Cedar Knoll throw of the fault has increased to 1,200–1,400 feet. Beyond this point, regional northward dip carries the resistant Kaibab beds below the much less resistant Moenkopi, and erosion has obliterated topographic expression of the fault.

Because the Sevier fault is nowhere in contact with lavas of the Uinkaret field, their relative ages cannot be determined. At least 25 to 35 feet of the movement along the Toroweap fault at the junction of the two must have taken place at the same time as that along the Sevier fault. Consequently, displacement along this section of the Sevier fault must have antedated the Stage III-b eruptions. H. E.

Gregory and N. C. Williams reported (personal communication) that in southern Utah the Sevier formation, of Miocene (?) age, is cut by the Sevier fault, but this does not fix the age of faulting in northern Arizona.

HURRICANE FAULT

Powell (1875, p. 186) first reported the Hurricane fault (Pl. 4) stating that the displacement is from 2,000 to 3,000 feet and that along this line of displacement is developed the Hurricane Ledge. Dutton concluded from detailed surveys of the fault (1882, p. 112–117) that at the Colorado River throw amounts to 1,500 feet or more and that immediately north of the river the fault breaks into several branches,

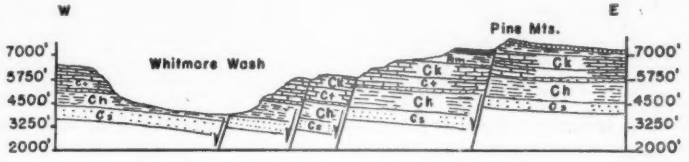

FIGURE 5.—*Hurricane fault west of Pine Mountains*

one of which cuts the old lavas of the Emma Platform. He also found that younger lavas crossed the fault line without displacement. Davis later re-examined this area (1903, p. 26–30); he stated that at Cold Spring, 2 miles west of Mt. Logan, Moenkopi and Kaibab beds are in contact at the fault, that the movements here exceeded 1,000 feet, and that old lavas passed unbroken across the fault.

Detailed studies of the area described by Davis show no basis for his conclusions. On the contrary, conditions are substantially as reported by Dutton. Immediately north of the Colorado River the fault divides (Pl. 1). The eastern branch, passing west of the Pine Mountains, cuts Stage I lavas underlying the Emma Platform (Fig. 5), with throw of 700 feet, then swings northward toward Mt. Logan (Pl. 4, fig. 3). Between the Emma and Logan platforms the fault line is buried by Stage III-a lavas, originating from cones 56, 57, 58, 64, 71, and 72; but a low scarp, probably marking the continuation of this fault, is clearly discernible. This scarp dies out in the Moenkopi which outcrops at the west base of Mt. Logan.

A second northeastward-trending branch of the main fault appears 5 miles north of the Colorado River with a displacement of 800 feet which diminishes northward. It has, in turn, a branch, parallel to the main Hurricane fault for a short distance, with throw of 300 feet.

Farthest west both Stage III-b and III-a flows bury the main Hurricane fault (Pl. 4, fig. 2). North of these lavas only the Moenkopi is exposed along the fault line, and the displacement decreases northward. At Cold Spring the throw is only about 700 feet (Fig. 6). Stage I lavas, approximately 100 feet thick are found west of the fault at Cold Spring, but none were found to the east except at Mt. Logan. The Cold Spring lavas may be a continuation of the Logan flows. Not only are the lavas displaced to the full extent of the throw measured in the sedimentary rocks, but they dip 8°–10° E. toward the fault, as do the underlying strata. This eastward dip indicates that they could not have flowed westward across the fault line as shown by

Davis (1903, Fig. 8, p. 28). Three miles north of Cold Spring, along the main Hurricane fault, the throw has decreased to 300 feet, the minimum amount anywhere observed.

Where the road from the town of Mt. Trumbull (Bundyville) to Nixon Spring intersects the fault, two faults, each of 75 feet displacement, run eastward from the main fault. Because of limitations imposed by scale, only one is shown on the geo-

FIGURE 6.—*Hurricane fault west of Mt. Logan*

logic map (Pl. 1). Dips in the Kaibab limestone along these minor faults amount to 10°–18° SE. In the next 3 miles northward throw along the Hurricane fault increases to 1,500 feet.

Possibly Davis' report of lava which flowed westward across the fault was based on the observation of talus blocks of lava which here cover the Moenkopi formation. Though there is today no lava in place between the widely separated remnants at Mt. Logan and Cold Spring, the presence of these blocks in an intermediate position indicates that a lava cap formerly extended most of, if not all, the way west from Mt. Logan to the Hurricane fault to connect with the Stage I lavas at Cold Spring. Difference in elevation of the patches on either side of the fault line and strong dip of the western patches toward the fault show later faulting of this sheet. Observed throws along the Hurricane fault on the west side of this portion of the Uinkaret Plateau can thus be dated as post-Stage I, pre-stage III-a lava.

Gardner (1941, p. 251) dated the major movement along the Hurricane fault in the Hurricane-Toquerville district of southern Utah as post-Miocene (?). Because the fault scarp is continuous between the areas, it seems likely, although not necessary, that faulting in the Mt. Logan area is also post-Miocene (?). There is no evidence in this area of repeated faulting such as is described by Gardner and Johnson to the north, and Stage I eruptions may tentatively be dated as late Miocene or early Pliocene, but preceding the faulting.

Dutton first noted a persistent downward flexing of beds on the western, downthrown, side of the Hurricane fault as the fault line was approached, and a corresponding though less sharply defined eastward dip of the beds on the upthrown block (1882, p. 114–115). He interpreted this as a result from faulting of an earlier monocline, in which the displacement had been opposite to that of the faulting—that is, the eastern side had been downflexed (Fig. 7 A, B). Gardner (1941, p. 254-259) has argued that the conditions observed in Utah do not support such a conclusion. He suggests (Fig. 7 C) instead that "the east block may have remained stationary or have moved upward without having been deformed, while the strata just across the fault to the west sagged downward with the maximum drop right at the fault and with rapidly lessening displacement to the west." Certain features noted by the writer along the fault from the Colorado River northward for 30 miles or more

FIGURE 1. VIEW SOUTHWARD ACROSS GRAND CANYON
Showing Hurricane fault south of Colorado River. Eastern branch of fault in lower left of view. Western branch, or main fault, at lower right, in Queantoweap Valley (*see* Pl. 1).

FIGURE 2. VIEW NORTHWARD UP WHITMORE WASH
Main line of Hurricane fault passes from left of view through center distance. Branch faults cut Permian strata of upthrown block at right of center. Stage III-b lavas have flowed from right to left across faults (*see* Pl. 1).

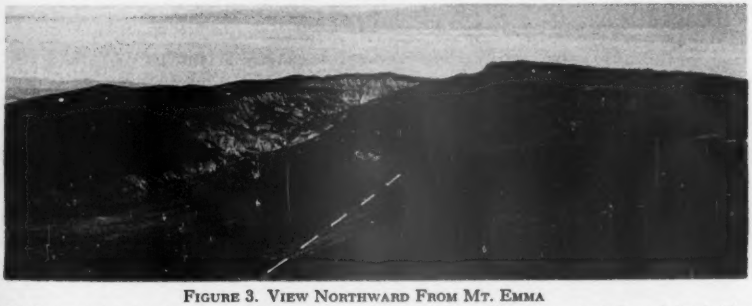

FIGURE 3. VIEW NORTHWARD FROM MT. EMMA
Mt. Logan right of center; Sawmill Mtns. left of center. Scarp of Hurricane fault branch buried by Stage III-a lavas indicated by dashed white line (*see* Pl. 1).

conflict with Gardner's hypothesis. At the river, strata on the downthrown block dip 15° E. On the upthrown block they dip 12° E. At Cold Spring, the Moenkopi and Stage I lavas dip 8°–10° E. on the downthrown block. Three miles east of the

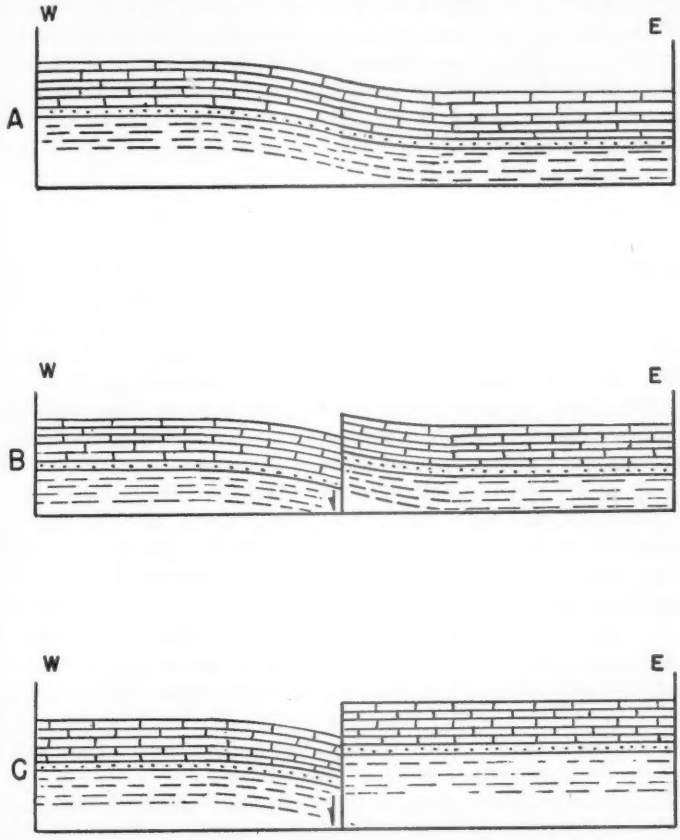

FIGURE 7.—*Schematic structure sections of Hurricane fault*

(A) Monoclinal flexing before faulting, as postulated by Dutton. (B) Present structure of Hurricane fault in Uinkaret region, after Dutton. (C) Structure of Hurricane fault in southern Utah, after Gardner.

fault the top of the Kaibab is nearly 1,000 feet lower than it is 3 miles west of Cold Spring on the downthrown block. Where the eastern branch of Hurricane fault cuts the Emma Platform, lavas on the upthrown block dip 10°–15° E. Northward from Cold Spring, downthrown beds dip 10°–15° E.; upthrown beds consistently dip 2°–6° E. These latter dips are not at the fault itself, but half a mile to a mile east, and hence need not indicate the maximum eastward dip on the upthrown side.

No explanation for these conditions is found in Gardner's suggestion whereas the solution advanced earlier by Dutton satisfactorily explains the relationships. The monocline may, however, have been limited to this part of the fault which is south of the area studied by Gardner, although Johnson (1909, p. 158) showed monoclinal

folding preceding faulting, in the reverse direction, in the area considered by Gardner. Gardner makes no reference to Johnson's work. Johnson described three successive periods of faulting along the Hurricane ledge but made no attempt to date them as exactly as Gardner has. Similar monoclines, preceding faulting and in the opposite direction, were described by Johnson (1909, p. 150) along the Sevier fault from Hack Canyon to Cedar Knoll where beds dip eastward on both sides of the fault, and along the West Kaibab fault by Dutton (1882, p. 128, 162) and Noble (1914, p. 76). If any downbending of the beds in the downthrown block in the manner Gardner describes occurred in addition to the ancient monoclinal flexing in the Cold Spring area, it was not detected by the writer.

MINOR STRUCTURES

Minor faults with displacements of 10 to 150 feet are common throughout the region. Five and a half miles west of Craigs Knoll, in the upper valley of Langes Run, is a small northward-trending normal fault with its eastern side upthrown 50 feet (Pl. 1). The downthrown beds dip 15° toward the fault but decrease to 4° a quarter of a mile west, while the upthrown beds show sharp drag in the immediate vicinity of the fault. It is not connected with any lavas or other later deposits, and it is thus impossible to determine its age. A mile and a half northwest of Craigs Knoll another small normal fault with throw of 10 feet cuts lavas of Stage III-a.

North and south of Schmutz Ranch, 4 miles east of Mt. Trumbull, are two small branches of the Toroweap fault (Pl. 1). The southernmost has displacement of 50 feet, and the beds in the downthrown block east of the fault are strongly tilted west. Undisturbed Stage III-a lavas cover the fault line. The northernmost branch is upthrown 50 feet to the east, and the beds between it and the Toroweap fault are tilted 9° E. The fault dies out within a mile to the northeast and does not come in contact with any lavas. Both blocks appear to have been tilted during faulting.

The saddle a mile southwest of cone 13 and 4 miles west of Vulcans Throne is bounded on the west by a normal fault, downthrown 240 feet to the east. Drag on the upthrown beds has caused them to dip 22° toward the fault. On the eastern side of the saddle is another normal fault, upthrown 120 feet to the east, with no drag. Stage III-b lavas immediately south of the displacements pass unbroken across the fault lines. A mile and a half west of this saddle is another normal fault, upthrown 320 feet to the east. Northeastward unfaulted III-b lavas bury this displacement. These three faults, very nearly parallel, are aligned in a northeasterly direction.

A normal fault extends from June Tank, 2 miles south of the Findlay Knolls, southeastward for 7 miles, the beds upthrown 100 to 150 feet to the east (Pl. 1). A broad lowland has been developed at the foot of the scarp, but no stream occupies the trough. Several small streams cross the fault without apparent regard for structure, suggesting that the fault may be relatively recent and that the streams are antecedent. Possibly also, before the overlying Moenkopi formation had been removed, extensive pediments, such as are common elsewhere on the Moenkopi, formed across the fault line. The streams were consequent on the eastward-sloping pediment surface and became superposed on the structure. Subsequent streams have since removed the weak Moenkopi along the lowland.

Faults are common throughout the Kanab Plateau. All are characterized by displacements generally less than 100 feet and rarely exceeding 150 feet, and all strike northwesterly or northeasterly.

Major joints of the Kaibab formation roughly parallel these directions, suggesting either that the minor faults developed along pre-existing lines of weakness defined by the joints, or that both owe their origin to the same set of stresses, perhaps applied during the deformation which resulted in the major faults and folds of the plateaus.

DESCRIPTIVE GEOMORPHOLOGY

INTRODUCTORY STATEMENT

The Uinkaret Plateau and neighboring Kanab and Shivwits plateaus lie in the Grand Canyon section of the Colorado Plateau province (Fenneman *et al.*, 1928). The region may be classified as block plateaus, because, while underlain by essentially horizontal strata, it is broken by three large and several small faults. Superimposed upon the plateau structure is the volcanic field.

Newberry noted (1861, p. 45–46) that throughout the region stream erosion has been the dominant process of dissection. Though some slight wind erosion may occur in the less resistant Moenkopi rocks in the north, the process is not common and has left no marks on the topography. No dune deposits have been found.

The Uinkaret Plateau as a whole is in the stage of late youth, although all the original surface has been destroyed. Stripped structural surfaces have been formed but are thoroughly dissected. In a superficial examination the observer is likely to be misled by the two distinct topographic textures. The most conspicuous is that of Grand Canyon and its larger tributaries. This coarse pattern contrasts strongly with the finer texture or smaller-scale dissection which characterizes the upland plateau surfaces where relief is generally 75 to 100 feet, and frequently more. Almost the entire land surface is in valley slopes, either of great magnitude in the Canyon, or of lesser magnitude in the smaller drainage systems which form a fine network over the surface of the plateau.

In the northern section where the weak Moenkopi outcrops, pedimentation is active. Large coalescent pediments are developed at the foot of Cedar Ridge and elsewhere along the Shinarump Bench. Smaller pediments at different elevations are common. Regrading of the pediments, as described by Johnson (1932), is an active process.

Dissection of the lavas has been very slight, but detailed examination shows that there are several stages. The older flows have been considerably eroded by streams flowing across or developing on their surfaces; the younger flows have been only slightly eroded, while the Stage IV flow is in its initial stage.

DRAINAGE DEVELOPMENT

Streams of the plateau fall into several genetic classes. Those which flow down Toroweap and Queantoweap valleys are subsequent where developed along zones of weakness produced by the Hurricane and Toroweap faulting or along weak-rock belts in the downthrown block, or consequent where flowing in the block basin formed

by faulting. There are several types of tributaries: those flowing down the escarpments of upthrown blocks or down back slopes of downthrown blocks are resequent; some, which have cut beyond the initial scarp, are obsequent. In both Toroweap

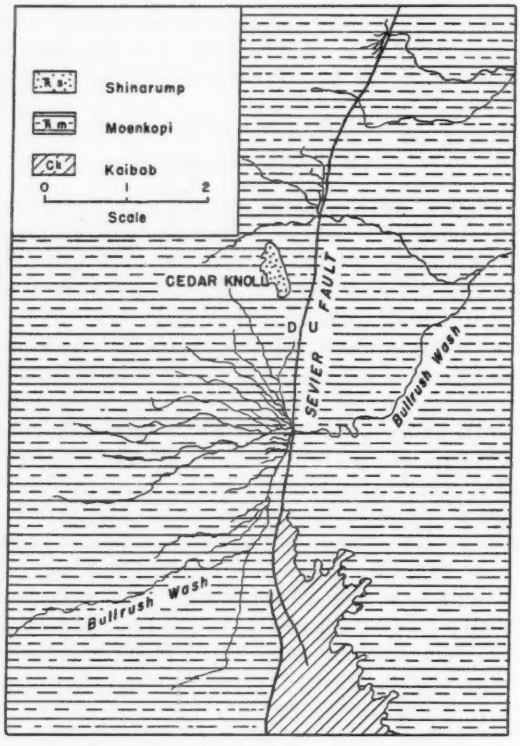

FIGURE 8.—*Map of Bullrush Wash at Sevier Fault*
Showing funneling of drainage at Sevier fault-line scarp near Cedar Knoll.

and Queantoweap valleys many consequent streams flow along margins or on the surfaces of the great lava cascades.

Langes Run, draining northward from Mt. Logan, follows with minor variations the regional northward dip of the plateau surface and may be classified as resequent. In places it follows flow margins and west of Craigs Knoll it follows a fault line for a short distance. Bullrush Wash, north of the Findlay Knolls, and Hack Wash in Hack Canyon both appear to be resequent with some subsequent portions. Both cross the Sevier fault; Hack Wash cuts through the escarpment and immediately enters a canyon 200 feet deep. Such development can best be explained by local superposition on pediment surfaces. Such pediments are now found immediately north of Cedar Knoll where weak Moenkopi beds are exposed on both sides of the fault line. If a similar pediment beveled the faulted mass farther south, as seems

probable, streams could cross the fault without deflection. When later erosion exposed the more resistant Kaibab below the Moenkopi, adjustment to structure would be rapid. At Bullrush Wash (Fig. 8) the Moenkopi has only recently been stripped from the upthrown block, and adjustment is but partially completed. One stream, more powerful than its neighbors, appears to have become deeply incised and to have diverted weaker neighboring streams which formerly flowed across the fault line. This has produced a curious funneling of drainage against the scarp, which is matched on a smaller scale 4 or 5 miles northward.

Because the strata dip gently northward away from Grand Canyon, there are two classes of streams developed on the plateau. The first flow northward, following in part the stripped structural surface of the resistant Kaibab formation, and have gentle gradients. The second consist of short streams with steep gradients, which flow southward into the Colorado River. The great contrast of slopes indicates that the divide between southern drainage to the Colorado and northern drainage to the Virgin River is migrating northward rapidly.

The Sawmill Mountain divide west of Mt. Logan is even more spectacular. Langes Run flows northward from this divide, at 7,200 feet, 60 miles to the Virgin River at 2,300 feet. South of the divide, formed on the Moenkopi, streams flow only 15 miles to the Colorado River at 1,600 feet. The divide is marked by a very strong southward slope which falls 1,500 feet in less than 2 miles. In contrast, the gentle northern slopes do not exceed 200 feet per mile. Such an asymmetrical and rapidly migrating divide may reasonably be explained by the former extension of the Logan lava cap across the area of the divide. The lava cup would have retarded divide migration until it was removed; then migration would have been greatly accelerated.

SLUMPING

Where the upper, resistant beds of the Kaibab formation have been cut through, exposing weak shales and gypsum beds lower in the formation, settling of overlying blocks is common and may amount to as much as 75 feet, though the average is nearer 25 to 30 feet. The phenomenon is well shown along the eastern wall of Toroweap Valley, near Hack Canyon, and northward along the Sevier fault line. Removal of the very weak shale and gypsum, in part at least by solution, allows the massive overlying joint blocks of limestone to drop downward from the main cap.

Around Mt. Trumbull, and to a lesser extent around the Emma and Logan platforms, large lava blocks have slumped. On the eastern face of Mt. Trumbull the lava cap is about 200 feet thick and lies on the Moenkopi formation, which is here composed of soft shales and marls, particularly weak when wet. The great weight of the lava cap has in places induced slumping of large blocks, some of them more than half a mile long and a quarter of a mile wide (Fig. 9). One such block, directly north of cone 107 on the eastern face of Mt. Trumbull (Pl. 1), has slumped so recently that a large undrained depression still exists between the back slope of the block and Mt. Trumbull. Along the northern face of Mt. Trumbull a shelf 100 to 200 yards wide is developed on the back slope of several tilted blocks (Fig. 10). This shelf lies some 500 feet below the summit, and the blocks have been rotated inward more than

40°. The slumping is old enough so that any initial closed depressions have been filled with alluvium. Similar shelves are developed at the base of the lava cap on Mt. Logan, though large-scale slumping has not been so extensive here nor in the old lavas of Mt. Emma; there are, however, a few small slump blocks along the western face of the platform at the Pine Mountains.

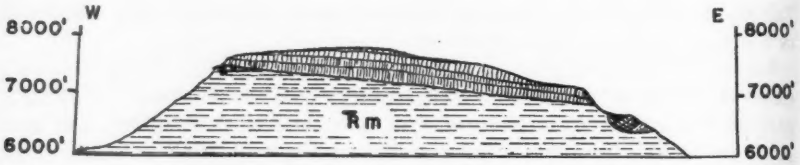

FIGURE 9.—*East-west structure section of Mt. Trumbull*
Showing slumping at eastern margin of basalt cap. Vertical scale twice horizontal.

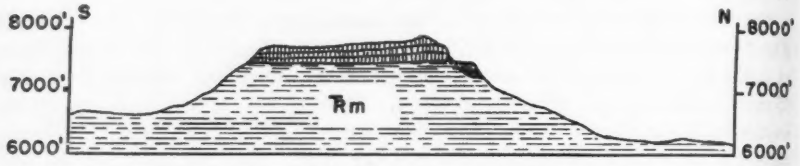

FIGURE 10.—*North-south structure section of Mt. Trumbull*
Showing slumping on north face. Vertical scale twice horizontal.

GROUND WATER AND EFFECTS OF SOLUTION

An arid climate combined with permeable rocks limits the groundwater supply in most of the area. However, important springs emerge from the relatively impermeable upper Moenkopi beds immediately below the lava caps of the Trumbull and Logan platforms. All may dry up during the more arid years except Big Spring, on the southern side of Mt. Logan, which has the largest catchment basin. Elsewhere ground water appears in small wet-weather seeps. In the Hermit shale along the Esplanade are a few springs, but the flow is small. All are located on the up-dip sides of tributary canyons. In addition to being limited in flow, springs along the Esplanade are unpalatable because of dissolved gypsum.

Sinkholes are common wherever Kaibab limestone is exposed but are usually small in the localities here studied. Some exceed diameters of 75 or 100 feet and have minimum depths of 25 feet, but the majority are 25 to 50 feet in diameter and less than 10 feet deep.

THE TRUMBULL, LOGAN, AND EMMA PLATFORMS

Dutton (1882, p. 103–104) noted Stage I lavas capping Mt. Trumbull, Mt. Logan, and Mt. Emma (Pl. 4, fig. 3). These he called "platforms," a well-chosen term because, on the relatively level surface of each, stand recent volcanic features. Lavas underlying the Trumbull platform vary from slightly over 500 feet thick on the northern side to 200 feet on the southeast. They are a series of thick flows, with well-

developed columnar jointing, separated by thin beds of volcanic ash and scoria. The crest of Mt. Trumbull rises 100 to 150 feet above the general level of the platform. In the north face of Trumbull a series of great columns cut across the beds of lava (Fig. 2). The writer interprets this as one of the necks up which the lava flowed during eruption. The fact that the flows thin and slope in all directions from this supposed neck supports this interpretation.

Lavas underlying the Logan platform are thinner than those of the Trumbull platform but are similar in every other way. They rest on the Moenkopi formation, present at the western end of the platform in very nearly its full thickness, and consist of several thick flows separated by thin beds of ash and scoria. All exhibit the well-developed columnar jointing characteristic of Stage I lavas. The platform slopes southeastward at a nearly uniform rate from a high point at its western edge, and the underlying Moenkopi wedge thins eastward. The eastern margin of the platform is buried by later lavas, and no Moenkopi is exposed.

The flows underlying the Emma Platform consist of several lava sheets similar in every way to those of the other two platforms. Along the western face the flows are 200 to 250 feet thick, from which point they slope eastward. An undetermined thickness of Moenkopi strata underlies the western edge of the platform. Eastward, the platform is almost completely masked by younger lavas, but on the eastern edge of the plateau the old lavas reappear from beneath those younger flows. Along the eastern margin the Moenkopi is very thin or absent.

A mile south of the Cove, 4½ miles west of Vulcans Throne, a very small remnant of Stage I basalt not recognized by Dutton was found. The lava, 130 feet thick, lies on Moenkopi, which is 200 feet thick at the western margin and 150 feet thick at the eastern.

Under each of the three larger platforms and the southern remnant, the Moenkopi thins eastward at a nearly uniform rate of 200 to 250 feet per mile. At the western tip of Mt. Trumbull 25 feet of Shinarump conglomerate is preserved in a wedge above the Moenkopi and disappears eastward as the sublava surface truncates successively older beds. As discussed later, these facts strongly suggest that the lavas poured out upon a pediment surface sloping southeastward and truncating the Moenkopi beds.

GEOMORPHIC HISTORY

PREVIOUS STUDIES

Powell (1875, p. 163) concluded that certain streams in the Grand Canyon region were antecedent. He held that, as a result of slow but long-continued uplift, immense amounts of rock had been eroded, and that carving of the canyons was an incident in erosion of the plateau region as a whole. He reported (1875, p. 198) that the lines of displacement "... have never determined the course of the streams.... All the facts concerning the relation of water ways of this region to the mountains, hills, canyons, and cliffs lead to the inevitable conclusion that the system of drainage was determined antecedent to the folding and faulting." According to Powell volcanic action was contemporaneous with the development of faults and folds and

continued until after erosion of Grand Canyon. He stated that a close relationship existed between lines of displacement and the location of volcanic vents, a correlation discussed elsewhere in the present paper. Powell further concluded that the climate had long been arid.

Dutton (1882, p. 219) reported that initial uplift of the Colorado Plateau region commenced in early Tertiary time and was followed by a long period of erosion in Miocene and Pliocene time, designated the "Great Denudation." By the close of this period the Grand Canyon region had been peneplaned (1882, p. 77, 119, 124), and on this surface were poured the first of the great lava flows (Dutton, 1882, p. 105, 107).

Extensive and rapid uplift took place in Pliocene time, and the great northward-trending fault systems developed. This uplift initiated a new erosion cycle in the Canyon, termed the Canyon cycle. Of the old erosion surface only lava-capped remnants are now preserved in the Trumbull, Logan, and Emma platforms. The Canyon cycle was thought to have been interrupted by a period of quiescence during which the Colorado widened its bed and cut the Esplanade (Dutton, 1882, p. 121, 226, 227), which is well developed in the Kanab and Uinkaret divisions. During subsequent uplift the present Inner Gorge was cut (Dutton, 1882, p. 94, 227).

Opposing Powell's view of aridity, Dutton (1882, p. 99, 194, 201) held that the climate during the "Great Denudation" was humid and that the region had suffered progressive desiccation since Miocene time. Evidence of increasing aridity was indicated by many large, untrenched valley floors above the level of the Colorado River, including Toroweap Valley east of the Kaibab Plateau. It was thought these were occupied by streams which disappeared during the Pliocene desiccation. Dutton further concluded from displacement of the Esplanade and other surfaces that the Colorado drainage was in the main antecedent to development of fold and faults.

Re-examination of Dutton's evidence led Davis to substantially the same conclusions concerning the Plateau and Canyon cycles. As additional evidence of the "Great Denudation," or Plateau Cycle, Davis (1901, p. 121–139) cited: (1) the mature valleys of the Kaibab and Coconino plateaus; (2) landslides of the Vermilion and Echo cliffs; (3) migration of divides near Pipe Spring and Cedar Ridge; and (4) bare scarps developed in weak Moenkopi shales. Davis (1901, p. 136–139), however, recognized that the great expanses of relatively flat surface resulted not from peneplanation but from stripping of weak overlying formations from underlying resistant beds.

Davis (1901, p. 143, 174), differing from Dutton on the age of the major faults and the history of the Canyon cutting, pointed out that the Esplanade was a structural surface, developed by stripping of weak members of the Hermit formation down to the resistant beds at the base, and hence had no cyclic significance. Consequently, displacement of the Esplanade at the foot of Toroweap Valley was not directly connected with faulting and could not be evidence of very recent movement. On the contrary, development of Toroweap Valley and Prospect Canyon along the fault line indicated to him considerable age for the fault. He further suggested (1901, p. 158–167) that the Colorado River was not antecedent to the structures of the plateaus but had developed in part as a consequent stream, following the irregular topography

which resulted from faulting and folding of the plateau, with some sections developing as a subsequent on belts of weak rock or along fault lines.

The valleys cited by Dutton as indicating an earlier, more humid climate were re-examined by Davis (1901, p. 187–192). He concluded that House Rock Valley was a normal subsequent valley, produced without regard to any major stream; that De Motte Park was similarly developed independent of any large, through-flowing stream; and that Toroweap Valley owed the elevation of its floor to lava which filled its mouth and retarded downcutting rather than to the disappearance of any large stream which had formerly occupied it. Davis also cited the presence of many small tributary canyons, which, though having smaller drainage areas than Toroweap Valley, had been cut to the level of the Colorado River. He therefore concluded that, in the absence of any conclusive evidence supporting Dutton's contentions, the climate had long been arid, though a slightly less arid period might have intervened during the Pleistocene.

Davis concluded that the Sevier and Hurricane faults were older than Dutton had supposed. Evidence for this included the great recession of the Vermilion Cliffs east of the fault as contrasted with that west of the fault at Pipe Springs (1901, p. 143–145; 1903, p. 12–15) and the fact that drainage from the downthrown block crosses the fault line to the upthrown block (1901, p. 145–146; 1903, p. 15–16).

Certain critical areas in the Grand Canyon region described by Davis were re-examined by Johnson (1909, p. 148–151) who came to the same conclusion respecting the great age of the Sevier fault. At Cedar Knoll erosion reversed the topography along the fault line, so that the escarpment at Cedar Knoll faces eastward. This reversal, obliteration of the fault line, and drainage from the downthrown to the upthrown block all indicated a relatively great age for the fault.

Johnson showed that the Toroweap fault was not identical with the Sevier fault, as Davis had supposed, but died out northward as a low monocline of decreasing displacement. Johnson (1909, p. 153–154) found as evidence of the great age of the Toroweap fault the broad flat-floored tributary valleys entering the main Toroweap Valley which must have developed contemporaneously with the main valley.

There has been wide disagreement among authors concerning the age of the Hurricane fault. Powell (1875, p. 186) stated that faulting followed development of the Colorado River. Dutton (1882, p. 112–117) stated that displacement took place before Stage I eruptions, and he concluded that faulting was relatively ancient. Huntington and Goldthwait (1904) described two periods of faulting in the Toquerville district separated by an erosion cycle. Johnson (1909) noted several periods of movement along the fault in southern Utah and northern Arizona but did not closely date these periods. Gardner (1941) postulated a somewhat similar history for this portion of the fault but made no statements concerning the time of faulting near the Colorado River.

AGE OF FAULTING

Movement along the Hurricane fault has already been shown to have occurred after extrusion of Stage I basalts and before Stage III-a eruptions. Presumably this movement took place at the same time as that dated by Gardner (1941, p. 251)

as post-Miocene (?) in the Hurricane-Toquerville district. The writer has shown that the Stage III flows of the Uinkaret field may be correlated with those of the San Francisco field. Sharp (1942, p. 502) states that Stage III flows of the latter field are post-Iowan-Wisconsin. Stage III flows of the Uinkaret field are probably of the same age, and the major displacement along Hurricane fault may hence be dated as post Miocene (?) but pre-Iowan-Wisconsin.

No evidence indicating the precise age of pre-lava movement along the Toroweap fault has been found in this area. The fact that the Toroweap and Sevier faults connect, and that the latter cuts sediments of Miocene (?) age in Utah, suggests that this early movement may have taken place during late Miocene or early Pliocene time. Displacement of III-a flows indicates post-Stage III-a faulting, the fault line being buried by Stage III-b lavas. Development of Toroweap and Prospect valleys along the Toroweap fault indicates that the first movement was fairly ancient, as deduced by Davis and Johnson. However, the statement by Johnson (1909, p. 154) that the broad mature tributary valleys to Toroweap indicate that both the main valley and the tributaries "... developed simultaneously after a period of base levelling" is not necessarily valid. That they developed concurrently cannot be disproved, but such development need not necessarily have followed base leveling of the region. Consequently, the long period of time which Johnson thought followed the faulting and during which the region was peneplaned is not necessarily required, although an ancient date for the first Toroweap faulting seems assured.

Absence of lava associated with the southern end of the Sevier fault makes it difficult to date this displacement. A mile and a half south of the junction of the Sevier and Toroweap faults the latter passes under undisturbed Stage III-b lavas, as already noted. As shown, some of the movement along the Sevier and Toroweap faults took place simultaneously in this section, and the Sevier fault may thus be dated as pre-Stage III-b in its southern part. Its post-Miocene age in Utah has already been noted. An obsequent fault-line scarp is present along the Sevier fault at Cedar Knoll. Similar drainage relations along the Sevier fault and Toroweap fault between the head of Toroweap Valley and the monocline indicate that the escarpment is of fault-line origin.

PAST CLIMATES

No new evidence on past climates in the region has been discovered in this area. Because of interior continental location it seems most logical to assume that the climate of the recent geological past was generally similar to that of today, except for a period of greater humidity during the Pleistocene.

CYCLIC DEVELOPMENT OF THE REGION

Introductory statement.—Weak Moenkopi shales under lava caps lying on well-developed erosion surfaces in the Trumbull, Logan, and Emma platforms indicated to both Dutton and Davis a long period of peneplanation prior to the canyon cycle. Both thought the great retreat of the higher terraces of the Plateau region as contrasted with that of the walls of Grand Canyon required two erosion cycles. In addition, Davis (9101, p. 120–136) cited the mature valleys of the Kaibab and Co-

conino plateaus, the migration of divides at Pipe Spring, landslides along Vermilion and Echo cliffs, and bare weak-rock slopes in the Pipe Spring-Uinkaret area in support of the hypothesis.

Pre-Stage I erosion surfaces.—Both Davis (1903) and Dutton (1882) repeat that the relatively smooth, level surfaces of Moenkopi underlying the lavas of Trumbull, Logan, and Emma platforms are peneplane remnants. Similar surfaces elsewhere in the Plateau province, either bare or covered by lava, are described (Robinson, 1907; Hack, 1942) as possible peneplane remnants. A peneplane developed in such weak rock as the Moenkopi would be expected to show low, uniform slopes with slight relief and with a deep residual soil cover. Along the major river valleys or flood plains, the surfaces should be horizontal and nearly plane, with tilted beds truncated by erosion.

In contrast, pediments developed in the Moenkopi commonly show very smooth, undissected, nearly plane surfaces, free of soil and with slopes of 10 feet per mile at the outer margin, increasing to 300 feet or more near the head of the pediments.

The surfaces beneath the Stage I flows of Mt. Trumbull, Mt. Logan, and Mt. Emma are nowhere exposed. Where the contact between the lavas and sediments was examined, there was no soil or gravel and no indication of extensive dissection. On the contrary, the contact is remarkably even. The slope, as determined from exposures around Mt. Logan, Mt. Trumbull, and Mt. Emma, is 250 to 300 feet per mile, increasing at the western tip of Mt. Trumbull. Such a surface could be a portion of a peneplane, but the uniformity of slope, absence of soil cover, and beveling of strata indicate the surfaces are more probably pediment remnants and not Plateau-cycle peneplane remnants.

Great retreat of high terraces.—Davis states (1901, p. 119), "It is evident that a broad denudation of the upper members of a stratified series can be contemporaneous with a narrow trenching of the lower members only if the former are relatively weak and the latter resistant." He concludes, without detailed discussion of the basis for his conclusion, that differences in resistance between the upper group of sediments, the Terrace series, and the lower group, the Canyon series, is insufficient to explain the difference in retreat of the terraces and canyon walls, and that an earlier cycle is required for the former.

Detailed comparison of the Canyon series with the Terrace series shows that the former (Pl. 3, fig. 3) consists of: the Tapeats, 75 feet of moderately resistant sandstone; the Bright Angel, shales and limestone 600 feet thick and of intermediate resistance; the Muav, limestone 470 feet thick and very resistant; the Temple Butte, limestone 335 feet thick, resistant; the Redwall, limestone 600 feet thick, very resistant; the Supai, 600 feet of sandstone, moderately resistant; and the lower 200 feet of Hermit, moderately resistant sandstone—a total of nearly 2900 feet. The resistance of the lower series compared to the upper is attested by: (1) the excessively steep walls of the lower series, frequently vertical or nearly so, and occasionally overhanging, instead of the sloping walls which even a new-cycle inner gorge would have in rocks of weak or moderate resistance; (2) the more moderate slopes commonly observed in the thick shale members of the upper series; (3) the well-known weakness of fissile shale as compared with sandstone and limestone when attacked by

weathering and running water. Above this group is 800 feet of moderately weak Hermit shale, then 1,100 feet of resistant Coconino sandstone and the Toroweap and Kaibab limestones. The Canyon series thus consists of about 4,000 feet of resistant rock with one intervening 800-foot series moderately weak rock. In the upper or Terrace series, above the Kaibab, is 1,600 feet of Moenkopi shale of very low resistance, 25 to 100 feet of resistant Shinarump conglomerate, followed by 600 feet of very weak Chinle shale. This total of 2,200 feet of very weak rock is then overlain by resistant Jurassic Glen Canyon sandstones which form the Vermilion and White cliffs. This great thickness of weak rock in the Terrace series, contrasted with only 800 feet in the Canyon series, would alone be sufficient to explain the greater retreat of the former.

Other factors may also have had their effect. Some southward thinning of the cliff-forming beds of the Terrace series is indicated by Baker, *et al.* (1936, p. 44–50), and under these conditions the terrace scarps would have retreated more rapidly than the canyon walls. The pre-Eocene (Wasatch) erosion surface, preserved farther north in Utah, may have extended southward and truncated the beds of the Terrace series (Fig. 11 A). If so, the Terrace scarps would have originated far north of the Canyon, as the beveled upper series was etched out by erosion. Moreover, the higher stratigraphic position of the Mesozoic beds would result in their early exposure to erosion and consequent greater retreat. The great retreat of the Terrace series as compared with the Canyon series does not, therefore, necessarily indicate a previous cycle of erosion.

Divide at Pipe Spring.—Davis cited (1901, p. 127–129) as evidence of two cycles the migration of the divide at Pipe Spring where two graded platforms head against each other, the eastern 200 or 300 feet below the western and separated by a distinct scarp (Fig. 12). After study it became clear that these two platforms are pediments, the western draining to Short Creek and the Virgin River, the eastern to Kanab Creek and the Colorado. Full explanation for the rapid encroachment of the eastern pediment on the western can be found in the advantage the former enjoys over the latter by virtue of the shorter distance to base level of the Kanab Creek drainage. Because individual pediments are not necessarily associated with any regional base level, regrading and encroachment of one pediment on another has no regional cyclic significance.

Landslides and mature valleys.—Davis (1901, p. 120–126) stated that landsliding along the Vermilion and Echo cliffs near Lees Ferry is evidence of a rejuvenation of the Grand Canyon region accompanying a second cycle of downcutting of the Colorado River. In the preceding "Great Denudation" the cliffs became "maturely stable," but following incision of the river, the weak Triassic shales were quickly eroded, causing cliff rejuvenation and landslides.

Strahler (1940, p. 298–299) cited the following weaknesses in Davis' hypothesis: (1) There is no youthful drainage system growing headward at the localities where landsliding ceases abruptly and the cliffs become stable. (2) Field study failed to substantiate Davis' statement that landslides are most abundant near the Colorado River where rejuvenation would be most active. (3) Davis based his arguments on the doubtful assumption that if but a single cycle of erosion were represented here it

FIGURE 11.—*Schematic north-south sections across Uinkaret region*

(A) Probable conditions at close of Eocene (Wasatch) deposition. (B) Probable conditions at time of Stage I lava eruptions. (C) Present geologic and topographic conditions.

would be in a late stage, and landslides, characteristic of early stages of a cycle, would not occur. Strahler explains the landslides as single-cycle erosion features resulting from oversteepening of cliff profiles by pedimentation. These conclusions are equally applicable to landslides along the Vermilion Cliffs immediately west of Pipe Spring, where regrading of the eastern pediment has exposed weak Chinle shales underlying the cliff-forming Navajo sandstone. This fully accounts for the slumping observed in the latter formation.

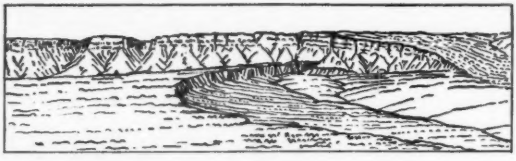

FIGURE 12.—*Migrating divide near Pipe Spring*

Vermilion Cliffs in background; broken by Sevier fault at far right. In center, Cedar Ridge, a stream-dissected escarpment between two pediments. Right-hand pediment, which drains into relatively near Colorado River, is being actively cut and is encroaching into higher pediment on left, which drains into Virgin River. (After a sketch by W.M. Davis, 1901.)

It is not within the scope of this paper to discuss the significance of the mature valleys of the Kaibab and Coconino plateaus, but field studies by Strahler (1944) and the author convinced both that these valleys are not, of themselves, indicative of a previous erosion cycle.

Absence of talus.—Bare scarps in Moenkopi beds west of Pipe Spring do not seem to indicate any significant change in the erosional history. Davis (1901, p. 133) explains the absence of talus at the base of the Navajo cliffs east of Gallup, New Mexico, as the result of " . . . the habit of massive sandstone to weather by crumbling, rather than by breaking into blocks." Since the Moenkopi and Chinle are here overlain by the same massive sandstone, previous cycle is not needed to explain a feature which is an expectable and normal result of having weak underlying beds capped by massive sandstones which weather to grains instead of blocks and consequently could supply no heavy talus cover to protect the lower slopes.

Conclusion.—Thus the evidence which Dutton and Davis thought indicative of peneplanation is inconclusive. None of the features cited are diagnostic of a previous cycle. All the known facts may fully be explained as a result of a single cycle of long-continued normal erosion in a semiarid region, accompanied by extensive pedimentation.

SUMMARY

The history of the Uinkaret region may be outlined as follows: At some time during the Tertiary, and before the Pliocene, the Colorado River had attained approximately its present course across the Plateau country, in part, at least, as a result of adjustment to crushed zones of faults and exposed belts of weak rock. By late Miocene or early Pliocene time it flowed at or near the level of the upper Kaibab beds. A large pediment or group of pediments then extended from near the river at least as far north as Mt. Trumbull. The thin wedge of Shinarump conglomerate preserved

beneath the northwestern tip of Mt. Trumbull may indicate that the earlier equivalent of the Shinarump Bench was located near Mt. Trumbull.

Onto this pediment surface, in late Miocene or early Pliocene time, flowed the Stage I lavas, from one or more vents. Sometime afterwards movement occurred along the Hurricane fault, and probably also on the Toroweap and Sevier faults. The displacement totaled more than 2,000 feet at the Colorado River, and the course of the river was maintained during this movement. The displacement may have been one of absolute elevation above sea level, with the Kaibab block raised the highest, and the western blocks uplifted progressively lesser amounts. But it is equally conceivable that the land first stood high, and that the blocks were dropped progressively nearer sea level to the west, the Kaibab block maintaining the highest elevation. Along the fault lines then formed the Hurricane Ledge, Queantoweap and Toroweap Valleys, and Prospect Canyon later developed. Concurrently, the cliffs of the terraces receded farther north, and the pediment surfaces were regraded. Remnants of these old pediment surfaces were, however, preserved beneath the lava caps of Mt. Trumbull, Mt. Logan, and Mt. Emma.

Sometime later, after the region had attained approximately its present configuration, the Stage II eruptions commenced. Perhaps by this time the river had cut to nearly its present depth. It has not been possible to date this period accurately. By the time Stage III eruptions began, the canyon had certainly been cut to its present depth, while along the Toroweap fault had been cut a tributary canyon reaching to river level, at least in its lower course (McKee and Schenk, 1942, p. 263–264). Stage II flows may have come down this canyon, though the lack of correlation between the Lower Canyon flows of the Inner Canyon lavas and any of the Uinkaret lavas makes it impossible to tell whether or not this occurred. During Stage II or early Stage III eruptions the Middle Canyon group of lavas reached the river, and dammed or ponded it for some distance upstream. That these flows were removed by the time Stage III-b eruptions occurred, is indicated by lava cascades which flow to the very bottom of the Inner Gorge. The last significant event in the history of the region was the eruption of a single Stage IV volcano, which took place so recently that no modification of the cone or flow is apparent.

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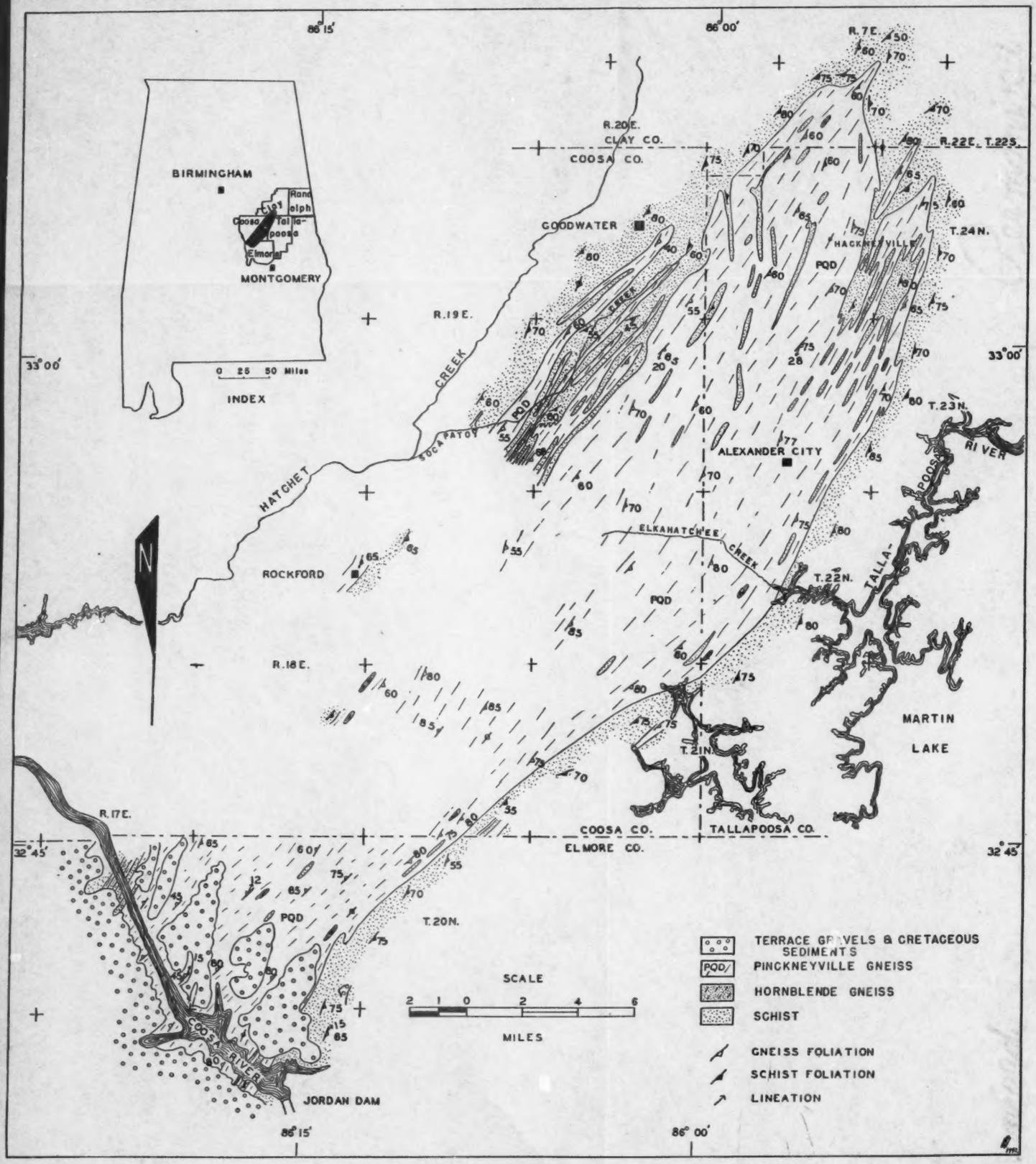

GEOLOGIC MAP OF THE PINCKNEYVILLE QUARTZ DIORITE COMPLEX

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PETROGRAPHY, STRUCTURES, AND PETROFABRICS OF THE PINCKNEYVILLE QUARTZ DIORITE, ALABAMA

BY HUGH RICHARD GAULT

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ABSTRACT

The Pinckneyville quartz diorite complex underlies an area in eastern Alabama extending from the Coosa River in northwest Elmore County northeast through Coosa and Talladega counties into Clay County.

Dark-gray, coarse-grained quartz diorite gneiss constitutes the major part of the complex, but there are smaller amounts of granodiorite and granite gneiss. Hornblende-biotite gneisses also occur and are considered to represent metamorphosed basic sills and dikes intruded before the quartz diorite complex. Schist septa lie within the complex and are usually parallel to the regional strike. The sequence of intrusion as interpreted from inclusions and petrographic data is (1) quartz

diorite, (2) granodiorite, (3) granite. Small but persistent amounts of allanite and epidote are characteristic of the rocks of the complex. The texture is subporphyritic and shows evidence of protoclastic deformation.

A northeast-southwest trend of inclusions, foliation, lineation, and contacts dominates. Primary flow structures are shown in the orientation of inclusions and micas. Secondary structures were superposed on the original flow structures shortly after intrusion and prior to complete consolidation. A prominent secondary structure is a lineation in *b*. Stretching in *b* is indicated by boudinage structures, the lineation, and small displacements of dikes. The intensity of the planar structures in the gneiss grades from strong to indistinct or lacking from east to west. Structures in schist inclusions indicate that some folding had taken place prior to intrusion.

Thin pegmatite, granodiorite, and granite dikes are common. Most of them are essentially parallel to the regional strike.

The petrographic boundary between gneiss and schist is always sharp. On a large scale, the gneiss structures are essentially parallel to the contacts, but slight to marked discordant relations occur at many localities. The schist structures usually parallel the contacts.

Petrofabric analysis confirms the structural field observations. The mica fabrics are *ac*-girdles and are most complete where the foliation is weakest. Microscopic *s*-planes do not always coincide with the observed foliation. The quartz fabrics are also *ac*-girdles but are not so well developed as the mica girdles. Quartz orientations cannot always be related to known *s*-planes.

It is concluded that the Pinckneyville quartz diorite complex represents a syntectonic intrusion—that is, that intrusion took place while external or regional forces were still active. The shape of the intrusion was controlled largely by the wall-rock structures which had developed earlier. The rocks show both igneous and metamorphic characteristics because the intrusion took place under regional metamorphic conditions.

INTRODUCTION

GENERAL REMARKS

Eastern Alabama is underlain by a large triangular area of metamorphic and igneous rocks which dip under the Coastal Plain sediments to the south and are the southern limit of the crystallines of the Appalachian system. Many of the problems of the Appalachian system to the north are also encountered in Alabama. The geological history of this southern limit is practically unknown, and only the most general structural and stratigraphic relations have been determined. Among these problems is the age and origin of the structures in some igneous rocks. Igneous intrusions are abundant in Alabama, and the Pinckneyville quartz diorite, originally described as a granite, is one of the largest.

The present study was undertaken to determine the origin of the structures and the time of intrusion of the Pinckneyville quartz diorite relative to wall-rock deformation. Detailed petrographic investigations were made for many of the rocks. This work is concerned principally with the intrusive mass, and actual age relations of the igneous and metamorphic rocks are not definitely determined.

LOCATION AND GEOGRAPHIC SETTING

The main body of the Pinckneyville complex (Fig. 1) strikes northeast in eastern Coosa and northwestern Tallapoosa counties. Its northern tip extends a few miles into Clay County. Thin sheets and small bodies of material described as similar to the Pinckneyville (Prouty, 1923a) extend northeast through Clay County into Randolph County. The southern boundary is at the Coosa River in northwestern Elmore County where the Coastal Plain sediments overlap the Piedmont. Only the main body of the complex, which covers an area approximately 40 miles long and 10 miles wide, is described in this report. The portion mapped in detail is shown on Plate 1.

The topography of the area is represented on the Talladega, Ashland, Wetumpka, and Dadeville quadrangle sheets. Physiographically the area lies in the Ashland Plateau district of the Piedmont Upland (Johnston, 1930). The altitudes of the hills range from 500 to 1000 feet, averaging 700–800 in the northern part and 500-550 in the southern part with a total relief of 50–250 feet.

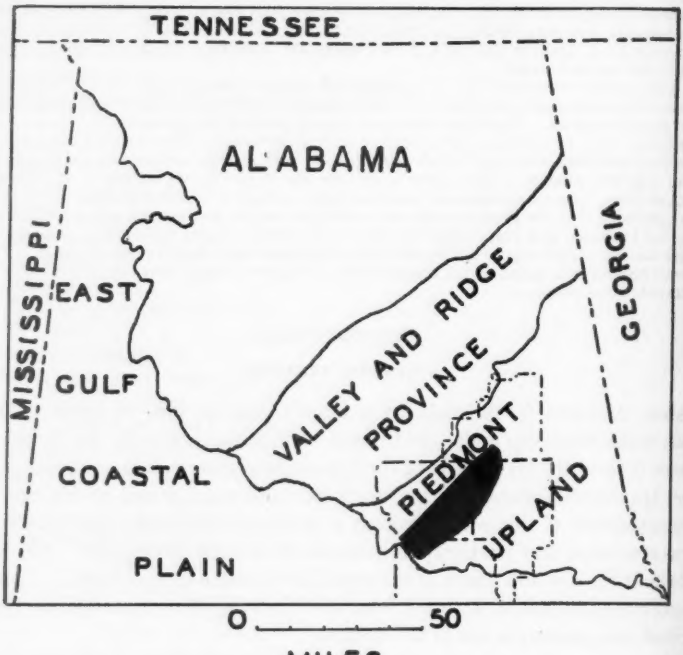

FIGURE 1.—*Location of Pinckneyville quartz diorite complex*
Physiographic provinces after Johnston (1930).

There is a faint ridge and valley topography in the northern quarter of the area formed by schist ridges flanking the lower, more rolling topography within the area underlain by the complex; the ridges become subdued or disappear to the south.

The area drains into Hatchet Creek to the west, the Tallapoosa River to the east, and the Coosa River to the south. The divide between Hatchet Creek and the Tallapoosa River runs roughly through the middle of the area to the Elmore County line. There are a number of tributaries that are fed by many smaller creeks.

Outcrops are moderately abundant near the main streams where the gradient of the tributaries is steep. Flat outcrops on hills and slopes are abundant in some areas. Most rocks are somewhat weathered, but quarrying and road maintenance indicate that relatively fresh igneous rocks can be obtained within a few feet of the outcrop surface. The metamorphic rocks generally are not eroded as rapidly as the igneous rocks, but they weather to greater depths. The soil cover is thin where

outcrops are flat; in other localities it is 30 or 40 feet thick, cut by large vertical-walled gullies. The ground-water level is 10 to 50 feet below the surface averaging 30 to 35 feet.

GENERAL GEOLOGY

The northeast trend of stratigraphic units and structures dominates the geology (Pl. 1). The foliation of gneiss and schist dips southeast except in Elmore County where the gneiss foliation dips northwest in the interior of the complex. Several miles east of the Pinckneyville complex in Elmore County the schist foliation also dips northwest.

The wall rocks are schist and phyllites with minor amounts of hornblende gneiss and other metamorphic rocks. They are included in the Wedowee and Ashland mica schist formations (Adams *et al.*, 1926; Stose, 1926), but no attempt is made to define stratigraphic units in this report.

Quartz diorite and granodiorite lenses of varying width and length intrude the schists and constitute the Pinckneyville complex. Schist septa of varying dimensions occur within the complex and are more abundant in the northern half. The rocks in the septa are the same types as those outside the complex. Schist structures make slight angles with the contact for short distances at many places, but, because of the general NE.-SW. trend of all structures, those in the schists essentially parallel the contact. In detail, discordant and disconformable structural relations between the gneiss and the schists are abundant, but, because the intrusion of the quartz diorite was controlled to a great extent by earlier formed structures in the schist, in general the structures of the gneiss are essentially parallel to the contact and to the schist structures.

PREVIOUS WORK

Practically all the previous work in the crystallines of Alabama has been reconnaissance. Tuomey (1858) first mentioned the Pinckneyville rocks; he spoke of a "whitish gneissoid granite" in the vicinity of Rockford and a fine exposure near Bradford, 6 miles south of Goodwater. He added that it belonged to the great gneissoid granitic belt which extends along the northern portion of the southern States. Tuomey seemed to have recognized the presence of schist septa (p. 78): "In hand specimen and even in the largest masses exposed, it presents no appearance of stratification; it is notwithstanding, interstratified with mica slates, has the same strike and dip and must therefore be called a gneissoid granite."

Smith (1875) mentioned the granite and gneiss in Coosa and Tallapoosa counties in and around Pinckneyville[1] and the alternation of gneiss and schist with occasional hornblende-bearing rocks. The gradation of granite into gneiss was also mentioned. Smith assumed that the granite and gneiss were pre-Cambrian.

Schmitz (1884) included the gneisses and granites with the Huronian and spoke of some granites resembling eruptive rocks.

[1] Pinckneyville, formerly a post-office, is now a cross-roads intersection. It is located in section 18, T. 22 S., R. 7 E., Clay County, just across the county line from Tallapoosa County.

Hitchcock (1885) published the results of a brief survey made between the Coosa River and the eastern boundary of Alabama for comparison with New England and Virginia to show that the crystalline belt did not stop in Georgia. He commented on the broad belt of gneiss between Pinckneyville and Alexander City as being reticulated in part with segregation veins similar to the Lake Winniscogee gneiss of New Hampshire. He also mentioned another oval portion of genuine granite, 6 miles long, similar to the Amoskeag quarry granite of Manchester, New Hampshire. Hitchcock concluded that the area of gneiss lay in a syncline. He considered that the gneisses were Huronian.

Until 1892, the entire region was called the crystalline belt of Archean age, and except for an occasional mention of granite or gneiss no other divisions were made. In 1892, there appeared the first of a series of articles by Phillips and in 1893 by Brewer on the gold deposits of Alabama. These articles dealt for the most part with the ore and mining operations. The origin of the gneisses was not regarded as having any practical significance in the development of the mines. Brewer (1896) seems to have recognized the alternating and generally concordant contacts of the igneous bodies as evidenced by this statement (p. 25):

"The structure is, I notice, generally conformable at the lines of demarcation between the gneiss, and either the slates or schists; and I failed to observe at any point any special change in the dip, or any appearance of greater deformation than is usually seen in this section of the Appalachian system .... that the gneiss will maintain its continuity longitudinally equally as persistently as do the mica schists, ..."

Smith (1896) spoke of the Alexander City gneiss belt as practically continuous from the Coosa River to the Georgia line. He recognized the continuous outcrop of this belt in Coosa County and the discontinuous outcrop north of Coosa County.

McCaskey (1908) describes a brief reconnaissance of several gold deposits in Alabama including the Hog Mountain deposit. He classified the rock at that place as granite but recognized its intrusive origin and that it showed only traces of dynamo metamorphism.

Aldrich (1909) dated the slates as Carboniferous. This is one of the first suggestions that all rocks in the region might not be pre-Cambrian.

Prouty (1923a) discussed the Pinckneyville more fully than had any previous worker. He gave the name Pinckneyville to the larger granitic mass in southern Clay County which represents the northern end of the main body. He also described a long, narrow belt of granitic material extending northeast in Clay County from Bluff Springs and considered it to be the same material as the Pinckneyville. Locally it was known as the Bluff Springs granite.

Although Prouty recognized four rock types of slightly different ages among the acid intrusives, petrographic studies were apparently insufficient to note the quartz diorite composition. The four types were: (1) biotite granite or the normal Pinckneyville, (2) smaller but regular granitoid strips near the borders of type (1), (3) irregular stringers and masses of light granitoid material in mica schists, and (4) pegmatitic material. He believed (3) was the oldest.

Prouty recognized that the Pinckneyville material intruded the schist and considered it genetically related to the Hillabee chlorite schist, an altered basic intrusion, and the acid rocks of the country.

Brown (1925) studied certain graphite deposits associated with small intrusives near Ashland, Clay County. He recognized three classes of intrusives: (1) gneissoid dikes approximately quartz diorite in composition, (2) aplites, and (3) pegmatites and quartz veins.

Adams (1930) also mentioned that strictly speaking the Pinckneyville was a quartz diorite rather than granite. He added that it was intruded into the schists as thick shoots and that the wall rocks influenced the development of the gneissic structure. In this report and in a later one (1933), Adams agreed with Prouty that the zone of intrusion of the Pinckneyville was along a fault zone.

Ellsworth (1934) studied the tin deposits of Alabama and agreed with the main points of earlier workers.

Park's (1935) investigation at Hog Mountain again revealed quartz diorite as the principal rock type. Hog Mountain is underlain by a small intrusive mass in Tallapoosa County, 4 miles east of the eastern edge of Plate 1 in sec. 15, T. 24 N., R. 22 E. It is considered to be an offshoot of the Pinckneyville complex to the west.

Park states (p. 5):

> "It was certainly intruded after most of the deformation that formed the Wedowee schists had taken place.
>
> "It is evident from the relatively poor schistosity and linear elements in the quartz diorite that the quartz diorite was intruded into the schist after the schistose structure was mostly formed. The quartz diorite cuts sharply across the schists planes in several places underground, although, in general, the contact parallels the schistosity. The presence of garnet and the development of strain shadows (wavy extinction) in the quartz suggests that the quartz diorite has undergone some regional deformation. In addition, the intrusive rock is sheared in places, especially in the part of the mass south of South Hill. It is thought that the quartz diorite was intruded along planes of weakness in the schist before the last crustal disturbance that formed the old folded Appalachian Mountains had entirely ceased. The quartz diorite, after solidifying, was subjected to a relatively minor amount of deformation and, being more competent than the schist yielded by clean fracturing under conditions where the schist yielded by flowing. . . .
>
> "The metamorphic effects at the contact of the schist and quartz diorite are slight."

PRESENT WORK

The present investigation represents 8 months of field work during the summers of 1940 and 1941 and 10 months of laboratory work during 1941 and 1942. The field work consisted of detailed structural studies and petrographic collections. Approximately 6,000 Brunton compass readings and many oriented specimens were taken. Because of the poor base maps available for Coosa, Tallapoosa, and Clay counties, mapping on a scale larger than 1 inch to the mile is difficult without plane-table control. Aerial photos are now available for this area but were lacking when the work was begun. Plane-table surveys were made at two strategic points, one on each side of the area (Pls. 4, 5). Pace and compass traverses were made everywhere except in Elmore County. Advance sheets for the Elmore County soil map (to be published) were used in that county, the scale being 2 inches to the mile, with only roads and drainage shown.

In the laboratory, 135 thin sections were studied, 19 of which were used in the petrofabric analysis. Mineral determinations and indices of refraction were made by the oil-immersion method. Modal analyses were made with an integrating stage.

ACKNOWLEDGMENTS

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PETROGRAPHY OF THE REGION

SEDIMENTARY ROCKS

Since the main emphasis is placed on the igneous rocks, the sediments and metamorphic rocks are described only briefly to present a general picture.

Capping the hills near the Coosa River in Elmore and Coosa counties are river-terrace gravels and the Tuscaloosa sandstone (Cretaceous). The gravels consist of coarse rounded quartz pebbles and sand. They show a rude stratification with thin intercalated lenses of sand and silt. They are dark brown and locally indurated with limonite.

The Tuscaloosa sandstone is usually fine-grained, but iron-cemented gravels finer than the terrace deposits crop out locally or occur as float. The sands are variegated light pink, reddish, yellow, buff, and white. Flakes of muscovite are common. Bedding is poorly developed.

Alluvial material is found along all the creeks and branches in the valley flats. The material is usually sandy and makes the best farming land of the area.

METAMORPHIC ROCKS

Mineralogy.—Biotite, muscovite, chlorite, and quartz are the abundant constituents with garnet, feldspar, pyrite, and epidote common but in small amounts. Graphite occurs sporadically in varying quantities. Sillimanite was found at one locality but may be widespread. Zircon, apatite, and sphene are common accessories, and allanite is present in some rocks. The opaque minerals include rutile, magnetite, hematite, limonite, leucoxene, and ilmenite (?). Tourmaline is usually an accessory, but tourmaline schists occur in the cassiterite-bearing pegmatite belt between Goodwater and Rockford.

BIOTITE: Biotite is persistent in all rocks and ranges from 50 per cent to a trace. Biotite flakes are usually small to medium lath-shaped grains (Fig. 2). In quartz-rich schists the grains are ragged and small. The flakes are often crinkled and folded and are occasionally sheared. Small zircon grains are the only common inclusions, and they are surrounded by pleochroic haloes. Sphene and black opaque material are included along grain boundaries or cleavage traces where several biotite flakes grow in a patch. Biotite is generally well oriented in one direction, but certain very coarse schists and strongly crumpled rocks have a poorly defined orientation.

Biotite is usually brown but where associated with chlorite has a slight green tint. 2V(−) varies from small to about 10°. The mean refractive index of cleavage flakes, determined for material from 12 localities, varies from 1.632 to 1.656, ± .002.

MUSCOVITE: Muscovite is found in almost all rocks and occurs as (1) sericitic material, (2) medium laths, and (3) porphyroblasts. The sericitic material grows in aggregates and patches and is the least abundant variety.

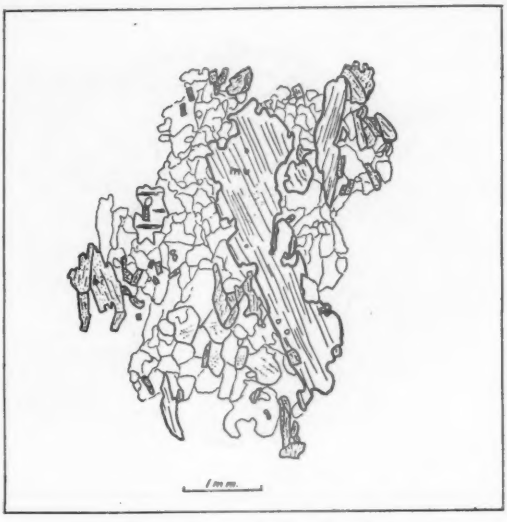


FIGURE 2.—*Quartz-biotite schist*

With muscovite porphyroblasts; sec. 17, T. 23 N., R. 20 E., Coosa County. (q) quartz, (mu) muscovite, (bi) biotite.

The second variety forms medium-sized laths usually larger than the biotite, but in some rocks it is small and ragged. Muscovite flakes are also crumpled, bent, and sheared. This variety more or less parallels the rock cleavage, but occasional grains cut across or are at right angles. Inclusions of dirty opaque material are more frequent than in biotite. Muscovite intergrows with biotite and may include small biotite flakes parallel to the cleavage traces.

The porphyroblasts of muscovite are of two types. The first is common and abundant where present. The crystals are 5 to 10 millimeters long and 3 to 5 millimeters thick. They have usually forced aside the earlier minerals and in some rocks show straining and bending. In such rocks the porphyroblasts try to assume the orientation of the other micas (D48). The second type is poikiloblastic and less common. The crystals cut the rock cleavage (Fig. 2) without forcing the other minerals aside. Some are strained.

2V(−) is 30°–35°. The mean refractive index of cleavage flakes, determined for material from eight localities, varies from 1.598 to 1.609, ± .002.

CHLORITE: Chlorite is abundant in some rocks almost to the exclusion of biotite, but elsewhere it is minor or lacking. The chlorite is commonly ragged and fuzzy, but where intergrown with muscovite or biotite it takes on an imperfect lath shape (Fig. 3). Chlorite is a medium green and has a slight brown tinge where associated with biotite. 2V(−) ranges between zero and 10°. The birefringence is low, and the chlorite appears isotropic or faintly blue under crossed nicols. The mean refractive index of cleavage flakes varies from 1.630 to 1.635. Optical properties suggest the aphrosiderite variety (Winchell, 1933).

FELDSPAR: Feldspar is common but not abundant. Grain sizes are 0.1–0.5 millimeter, about the same as that of the associated minerals, and the grains are more or less equidimensional. Oligoclase is most abundant with lesser amounts of albite and andesine. Potash feldspar is rare or absent.

GARNET: Porphyroblasts of garnet occur in many rocks. The crystals are 1 millimeter to more than 1 centimeter in diameter. Where garnet has good euhedral outline, the surrounding minerals wrap around the porphyroblast as if forced aside by growth of the garnet, and the curved arrangement of inclusions within the crystal suggests that the garnet was rotated during its growth (Fig. 4).

Anhedral garnet does not force aside adjacent minerals noticeably and includes quartz, epidote,

FIGURE 3.—*Crumpled muscovite-chlorite schist*

Sec. 1, T. 24 N., R. 21 E., Tallapoosa County. (chl) chlorite, (mu) muscovite, (py) pyrite.

and other minerals poikiloblastically. Earlier structures may be continuous through these anhedral garnets. In some crumpled schists and quartz-rich rocks, the garnets are fractured and dragged out into the foliation.

ACCESSORIES: Pyrite is always anhedral in irregular, elongate patches. Epidote occurs as small anhedral or short columnar grains. It is a faint to medium yellow green. Very small zircon crystals occur in biotite surrounded by pleochroic haloes. Larger, more rounded zircon grains and tourmaline are scattered through the rocks similar to heavy minerals in sediments. The tourmaline schists contain long, slender, black needles in the cleavage planes.

Sillimanite is of two types. The first occurs in bundles of thin fibers of light dirty brown. The second type, which occurs with the first, is long prismatic crystals with cross fractures and diamond-shaped cross section.

Rocks.—The metamorphic rocks include schists and phyllites with subordinate amounts of quartzitic and epidote-quartz rocks. Hornblende schists and gneisses also occur and are discussed with the igneous rocks.

Where muscovite predominates, the rocks range from fine-grained phyllites to coarse-grained schists. They are tan, brown, dark gray, silver gray, or silver green. With increase in biotite they become darker. As a rule, biotite schists are finer-grained than muscovite schists.

The chloritic schists (Fig. 3) are dark green brown, and many are coarse-grained. The quartz schists carry varying amounts of muscovite and biotite, are fine-grained,

and light or dark gray (Fig. 2). The graphite schists are fine-to medium-grained dark rocks with abundant muscovite and quartz. Pyrite is common in the graphite schists, and on weathered surfaces the rocks are coated with an efflorescence of white alumlike material. Sillimanite schist was noted at one locality, and it is a coarse-

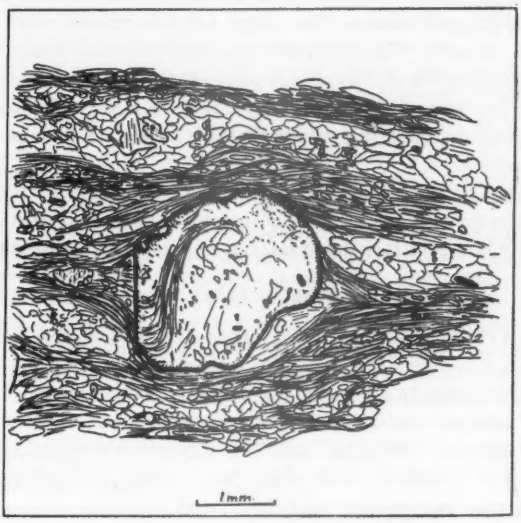


FIGURE 4.—*Quartz-muscovite schist*

With porphyroblast of garnet; Elkahatchee Creek, Tallapoosa County. (ga) garnet, (mu) muscovite, (q) quartz, (py) pyrite, (p) plagioclase.

grained, dark-brown biotite-sillimanite schist. Prouty (1923a) and Brown (1925) report sillimanite in Clay County which suggests that the occurrence is probably more widespread than indicated here.

Garnet and muscovite porphyroblasts are common and are more abundant in the mica schists than in the quartz schists (Figs. 3, 4). Crumpling and folding of mica plates and cleavage planes is common but does not occur everywhere. Crumpling and folding of the cleavage is more prominent in the mica schists, particularly the coarse-grained types. A fine banding is visible in many specimens and is due to varying proportions of quartz and mica.

Quartzitic and epidote-quartz rocks are widely distributed but quantitatively minor. The quartzitic rocks contain slight amounts of garnet, epidote, and at a few localities traces of hornblende. They are dark green to black or dark gray with spots of pink garnet. The epidote-quartz rocks have essentially the same mineralogy, but epidote is much more abundant. They are lighter in color than the quartzitic rocks.

Modes of some metamorphic rocks are given in Table 1. This table does not include any muscovite schists.

All varieties of metamorphic rocks are uniformly distributed, but graphitic schists

are more abundant on the western side of the area. The schists are intercalated with one another, and no stratigraphic units are defined.

Regional thermal metamorphism due to the intrusion of the Pinckneyville complex was not observed, either because there is none in these highly metamorphosed schists or the author's observations do not extend far enough out from the intrusive. Locally there is a slight coarsening of the schists, but fine-grained rocks are also found adjacent to the gneisses. The quartzitic and quartz-epidote rocks may represent contact metamorphic types of rocks developed near the contact or in the septa within the complex. The development of sillimanite may be a local phenomenon, but the reported widespread distribution suggests that it is a regional mineral due to regional metamorphism.

The minerals and petrographic types of the metamorphic rocks within the complex and on the borders are of the same grade of metamorphism for at least a mile from the main gneiss body and probably much farther, at least toward the east.

IGNEOUS ROCKS

Mineralogy.—Feldspar, quartz, biotite, muscovite, and hornblende are the principal rock-forming minerals. Plagioclase, principally oligoclase and andesine but ranging from labradorite to albite, is the most abundant feldspar. The grains are large, most of them are twinned and many are zoned. Inclusions of epidote and muscovite are common. Orthoclase and microcline occur in about equal proportions but are much less abundant than plagioclase. Some myrmekite is recognized. Most feldspar grains show evidence of deformation.

Quartz generally occurs as large patches of several grains. Many grains show undulatory extinction. Biotite laths are intergrown in patches many of which contain epidote nuclei. Several types of muscovite are recognized. Biotite and muscovite show strain shadows. Hornblende crystals are stumpy prismatic grains.

Epidote and allanite are minor but persistent minerals. Sphene, pyrite, and apatite are common accessories. Zircon, garnet, and tourmaline are sporadic. Chlorite and magnetite are alteration products. Other minerals are carbonate, sericite, hematite, limonite, and leucoxene.

FELDSPARS: Plagioclase is the most common and abundant feldspar and occurs in all rocks except a few pegmatites and some altered basic rocks. Andesine and oligoclase are most abundant, but the plagioclase ranges from labradorite (Ab_{46}) to albite (Ab_{98}). The blocky subhedral to anhedral grains suggest phenocrysts. They are surrounded by mica and quartz and smaller feldspar grains (Pl. 2). The plagioclase occurs in augen which are more pronounced in the more strongly foliated rocks. In faintly foliated or slightly lineated rocks, the augenlike texture is only suggested.

Grain boundaries are either sharp and distinct or indefinite and irregular. The sharp boundaries are outlined by micas, large patches of quartz, or fine-grained accumulations of quartz and feldspar. Contacts with mica flakes are sharp and straight where mica cleavage parallels a crystallographic direction in the feldspar and jagged where muscovite or biotite protrudes into the feldspar. Figure 7a shows the differential growth of plagioclase twin lamellae to produce rectangular ridges. Biotite and quartz grow on these ridges.

In some rocks muscovite laths wrap around the corners of plagioclase or are more or less parallel to the outlines of the grain. Muscovite is also intergrown with plagioclase with a texture similar to myrmekite.

The boundaries of granulated plagioclase grains are irregular and indefinite. Where two large

feldspars are in contact their boundaries are angularly irregular and interlocking. In many cases, two large feldspars are separated by a thin seam of muscovite. The small plagioclase grains are anhedral but more or less equidimensional. Their outlines are not emphasized by micas. Many plagioclase grains are zoned. Normal zoning is the usual type with an average difference in composition of 12 per cent (Ab) between core and periphery. Zoning is only faintly visible in albite. Recurrent zoning occurs rarely, and invariably the outer edges are more sodic than the interior. In such cases the core may or may not be more calcic than the intermediate zones. In many feldspars zoning is restricted to portions of a crystal. In some rocks the plagioclase has a more sodic border which is not due to zoning but seems to be a later overgrowth.

Twinning is usually according to the albite law, but pericline and Carlsbad twins have been observed. All three occur together in some grains. Albite twinning is more distinct in the more calcic varieties. Commonly one grain shows several roughly rectangular areas of twinned material, and the lamellae of one area make angles up to 90° with those in another area. This development of areas with their twin lamellae at angles is apparently due to protoclastic deformation (Fig. 5a). The twin lamellae vary in width and terminate abruptly, the terminations in some crystals paralleling possible crystal faces (Fig. 5b).

Inclusions of mica laths and small epidote crystals are common and abundant in many plagioclase grains. The small mica and epidote inclusions are oriented parallel to crystallographic directions (Fig. 6). The epidote is not so well oriented as the mica. The small inclusions are not always uniformly distributed throughout the grains but are concentrated either in the interior or in certain restricted portions of the grains. They seem to be more common in the calcic portions. The pattern is similar to that described by Ingerson (1938) and Cloos and Hietanen (1941). Large biotite and muscovite flakes, quartz, and hornblende also appear to be included at random in a few crystals.

Megascopically the plagioclase is white to glassy. Under the microscope it is fresh except along certain fractures where it appears cloudy because of bleaching.

Orthoclase and microcline replace plagioclase with the concomitant development of myrmekite and quartz (Fig. 5c). Replacement of large grains is rarely complete. Myrmekite is always accompanied by potash feldspar. Perthites were not observed.

Microcline and orthoclase occur in about equal proportions, although in any one rock there may be an excess of one. The mode of occurrence of each is the same except where microcline replaces plagioclase. In general potash feldspar is interstitial to all minerals except quartz. Much of it occurs in fine granular areas with quartz, and some with small plagioclase grains. The grain boundaries are very irregular and indefinite in these areas.

Where potash feldspar exceeds 10 per cent, its habit is more perfect, and the grains are larger, but some still occurs in the granular areas. In pegmatites, orthoclase and microcline crystals are as large or larger than the associated plagioclase grains. Potash feldspar increases with an increase in albite in the plagioclase.

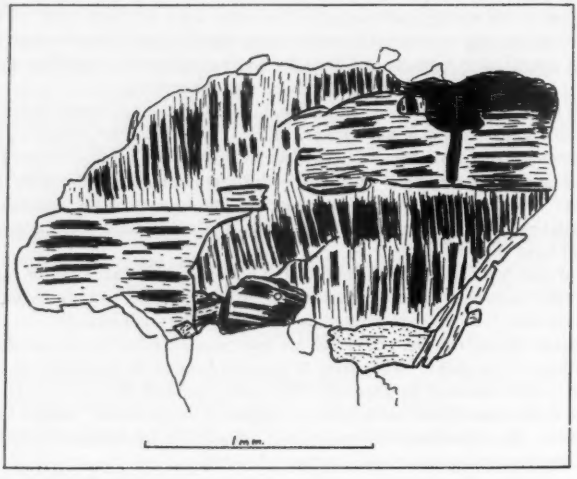
Inclusions in the potash feldspars are rare but where present are commonly muscovite. The feldspars weather and are cloudy. Both potash feldspars are white, pink, or buff.

In most rocks feldspar is deformed. The corners of the crystals are broken off, and the edges are granulated. The fragments are scattered in the adjacent finer-grained material. Where feldspar grains adjoin small shear zones, the boundaries of the grains are sharp and usually straight regardless of their crystallographic directions. Bent twin lamellae and undulatory extinction across the lamellae occur in some crystals.

QUARTZ: This mineral is second to feldspar in amount. It is clear with few dustlike inclusions in contrast to the quartz grains in some schists which show inclusions with trends parallel to rock structures. In the coarse-grained rocks the quartz is in large patches made up of several grains and fills in around other minerals. The size of the quartz grains varies greatly, but the large patches rarely equal the size of plagioclase grains. Within the patches the individual grains have sutured, irregular, or granitoid boundaries. Many of the grains within one patch are distinct units with different orientations. Flamboyant quartz is common. In some large grains the undulatory extinction swings across jerkily, suggesting that the whole grain is composed of smaller pieces, whereas in other cases the extinction swings across several grains uniformly without regard for grain boundaries.

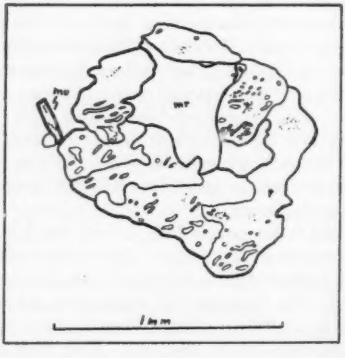
Short narrow zones of quartz lie between large crystals of plagioclase, and the zones are comprised of needlelike strongly strained grains which follow the outline of the feldspar grain.

BIOTITE: Biotite is common in all rocks except pegmatites and some dike rocks. The biotite grains are blocky laths much smaller than the plagioclase grains. They are generally dark brown



a

b



c

FIGURE 5.—*Structures in feldspar*

a. Fractured and strained plagioclase; sec. 1, T. 23 N., R. 20 E., Coosa County (crossed nicols)
b. Twinning in plagioclase; sec. 20, T. 24 N., R. 21 E., Tallapoosa County (crossed nicols)
c. Myrmekite intergrowth of quartz and plagioclase; sec. 24, T. 20 N., R. 18 E., Elmore County. (p) plagioclase, (mr) microcline, (mu) muscovite.

but also dark green brown, light brown, and warm red brown. Color differences may be due to weathering or hydrothermal alteration, but there is no apparent relation between the color and the composition of the biotite, the rock type, or the associated minerals.

The mean index, measured on cleavage flakes from 17 localities and 5 rock types, ranges from 1.623 to 1.657 ± .002. 2V(−) varies from zero to about 10°. Pleochroism is light yellow to brown. A

relation between index and composition and rock type is suggested but is not always clear. The biotite associated with hornblende has the lowest indices, as might be expected (Tsuboi, 1936). The mean index of biotite in the granitic rocks seems to increase slightly with an increase in potash feldspar. This distinction is more apparent between the quartz diorite and granodiorite than between granodiorite and granite. Correspondingly there is a greater petrographic difference between quartz diorite and granodiorite than between granodiorite and granite.

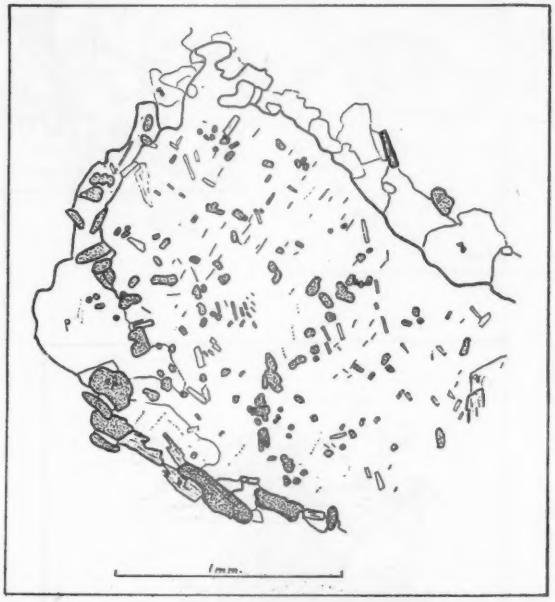


FIGURE 6.—*Inclusions of epidote and white mica in plagioclase*
Sec. 2, T. 21 N., R. 20 E., Coosa County. (p) plagioclase, (ep) epidote, (bi) biotite, (q) quartz.

Most biotite occurs in patches of intergrown, predominantly subparallel flakes. Some biotite flakes are attached to feldspar grains, and others occur as individual flakes. Small, ragged biotite flakes which appear to be shredded remnants of larger flakes occur in areas of fine-grained quartz and in small sheared zones. Epidote occupies the center of many biotite patches or is partially surrounded by biotite laths (Fig. 7c). The epidote seems to have served as a nucleus on which the biotite grew.

Biotite flakes usually show a slight undulatory extinction in all rocks and in some rocks are bent or sheared (Fig. 7b). Inclusions are commonly small zircon, allanite, or epidote crystals surrounded by pleochroic haloes (Fig. 9). Sphene crystals are locally included between grains and along cleavage traces.

Biotite is altered to a chlorite along many joints. Magnetite develops in the chlorite derived from alteration of biotite.

MUSCOVITE: This mineral occurs as (1) medium laths, (2) small inclusions in plagioclase, (3) feathery mica in plagioclase growing on muscovite (1), and (4) porphyroblasts.

Laths of muscovite (1) are most common and abundant, and their mode of occurrence is very similar to that of biotite. Many surround epidote grains or have grown on them. Muscovite (1) is in patches which are more elongate than those of biotite. These muscovite patches wrap around the plagioclase grains outlining the feldspar augen where they are well developed. Thin blades of muscovite lie between many large plagioclase crystals. Here and there the muscovite blades have

concussion marks near the corner of the plagioclase grain as if the plagioclase crystal had been ground into the muscovite. Within the muscovite patches are many small sheared zones, and the mica in the sheared zone is finely comminuted. A few large flakes appear to be included in plagioclase.

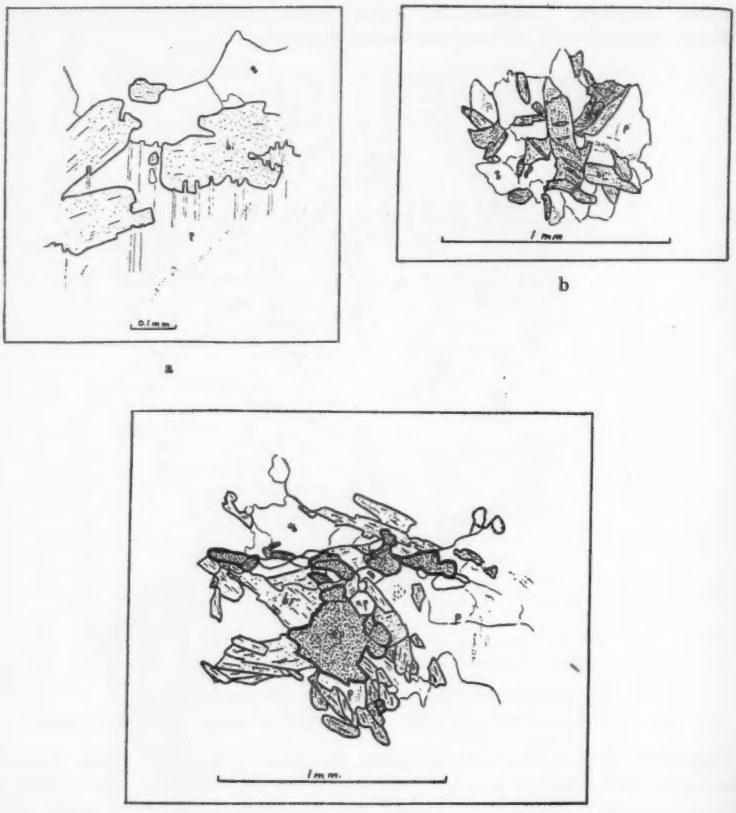


FIGURE 7.—*Structures of biotite*

a. Biotite growing on plagioclase twin lamellae; sec. 16, T. 20 N., R. 18 E., Elmore County
b. Bent and sheared biotite in an inclusion; sec. 14, T. 20 N., R. 13 E., Elmore County
c. Biotite growing on epidote; sec. 5, T. 19 N., R. 18 E., Elmore County. (bi) biotite, (ep) epidote, (ap) apatite, (p) plagioclase, (q) quartz.

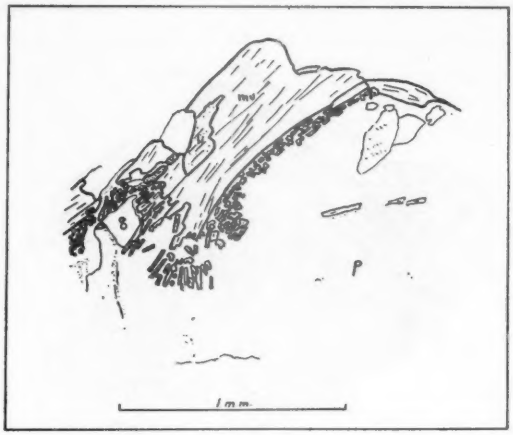
Muscovite (1) flakes are generally larger than biotite, show faint undulatory extinction, and are bent and faulted. The mean index of refraction determined on the cleavage flakes for material from 25 localities and representing all rock types ranges from 1.596 to 1.606 ± .002. 2V(−) is between 30° and 35°.

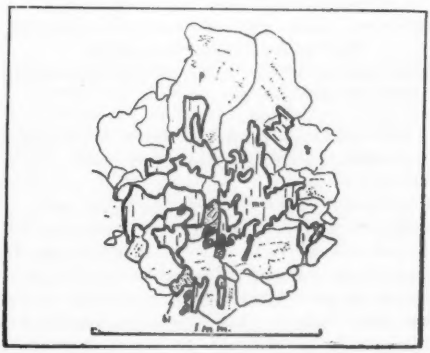
Small lathlike inclusions, muscovite (2), have already been discussed under inclusions in plagioclase.

Muscovite (3) is common in the oligoclase-bearing rocks and in many cases is associated with myrmekite, This type of muscovite is developed on muscovite (1) at its contact with feldspar and

within feldspar. The granodiorite at Rockford (Fig. 8a) contains an appreciable amount of muscovite (3).

Muscovite (4) is porphyroblastic and poikiloblastic (Fig. 8b). Many grains are slightly strained. This type is not abundant and occurs principally in those rocks containing abundant potash feldspar.





b

FIGURE 8.—*Muscovite*

a. Muscovite (3) and muscovite (1); Rockford, Coosa County.
b. Porphyroblastic muscovite enclosing quartz and other minerals; sec. 14, T. 20 N., R. 17 E., Elmore County. (mu) muscovite, (p) plagioclase, (q) quartz, (bi) biotite.

EPIDOTE AND ALLANITE: These two minerals usually do not occur in large quantities, but they are characteristic of the Pinckneyville complex. Epidote occurs in almost every rock ranging from a trace to 25 per cent, commonly from 1 to 6 per cent. Allanite is always surrounded by epidote.

There are three types of epidote: (1) euhedral and columnar crystals, (2) anhedral pistachio-green grains, and (3) small inclusions in plagioclase.

Epidote (1) is the most common type and is euhedral to subhedral and columnar. It is usually surrounded by biotite or muscovite (Fig. 7c). It always surrounds the allanite grains, but not every

epidote grain has an allanite center. Sphene and apatite are common associates of epidote within the mica patches. Epidote (1) is colorless or shows a very faint yellow pleochroism indicating a low iron content. Some of the material identified as epidote may be clinozoisite as a few grains give optically positive (?) interference figures. 2V is large in all grains.

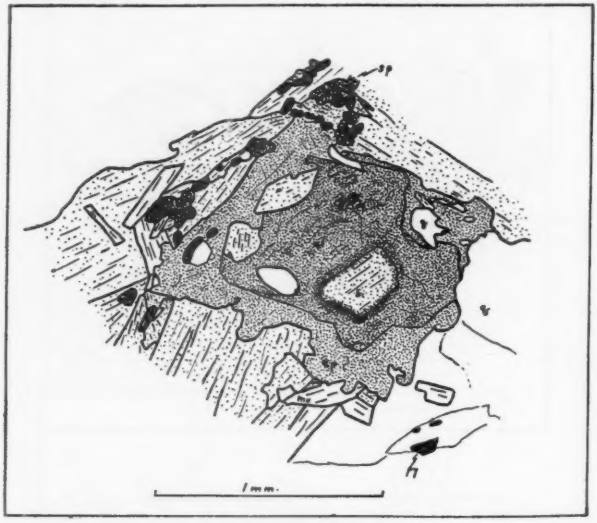


FIGURE 9.—*Allanite core in epidote*

Note pleochroic haloes around biotite; sec. 20, T. 24 N., R. 21 E., Tallapoosa County. (al) allanite, (ep) epidote, (sp) sphene, (q) quartz, (mu) muscovite, (py) pyrite.

Epidote (2) is usually anhedral, and its pistachio-green color contrasts with the faint yellow or colorless epidote (1). It occurs in epidote-rich rocks and in joints. Epidote (2) is not abundant. Both epidote (1) and (2) are granulated.

Epidote (3) occurs as inclusions in plagioclase and has been discussed.

Allanite is generally subhedral to euhedral and varies from minute grains to large crystals (Fig. 9). It is brown yellow to dark yellow brown. In some grains the color of the cores is lighter than the margins. A few crystals show twinning. The birefringence is weak to strong. Allanite is always surrounded by epidote, but the two minerals are not optically continuous.

A black mineral with a pitchy luster in a thin pegmatite seam is identified as allanite (Larsen and Berman, 1934). It is isotropic, greenish gray, and shows conchoidal fracture. The index of refraction is 1.680 ± 0.002.

HORNBLENDE: The hornblende crystals are stumpy prismatic grains with irregular subhedral outline. Grain sizes vary from 1 to 10 millimeters, and the larger grains are blocky. Optical properties indicate that the hornblende is pargasite (Larsen and Berman, 1934). The birefringence is 0.020 to 0.023 with beta ranging from 1.663 to 1.671. The pleochroism is yellow green to light yellow to dark green. $2V(-)$ is large, and $Z \wedge C = 23°$; $r > v$.

In rocks with abundant feldspar hornblende is usually associated with biotite and epidote, and the minerals are intergrown. Ordinarily the hornblende is clean and fresh. Inclusions in hornblende (sphene, biotite, magnetite, zircon, quartz, or feldspar) are not common. Quartz and feldspar are included poikiloblastically. The biotite inclusions are subhedral and seem to have mutually intergrown with the hornblende. Sphene inclusions are euhedral. Zircon inclusions are rare and are surrounded by pleochroic haloes. Muscovite does not occur with hornblende except as inclusions in the associated plagioclase grains.

Hornblende alters to a chlorite which differs from the chlorite associated with the biotite. In some rocks an intermediate stage of the alteration of hornblende to chlorite is recognized by a band of variable width of lighter green to colorless material which is developed on the outer margins of the hornblende crystals (Fig. 10). The indices of this outer band are lower than those of hornblende, but the birefringence is about the same. The altered borders are in optical continuity with the unaltered hornblende.

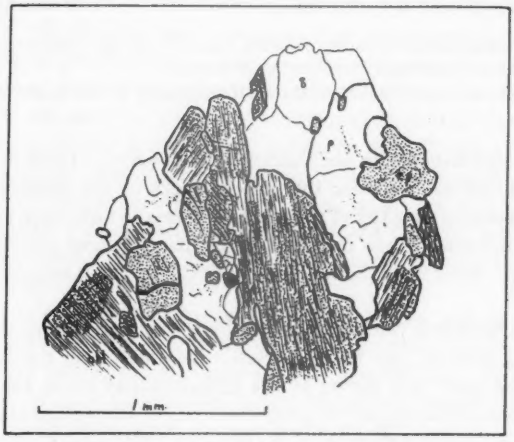


FIGURE 10.—*Hornblende altering to chlorite*

Sec. 8, T. 22 S., R. 7 E., Clay County. (hbn) hornblende, (chl) chlorite, (q) quartz, (p) plagioclase, (ep) epidote.

CHLORITE: Two varieties of chlorite are identified; one is associated with biotite, and the other occurs in or near the hornblende-bearing rocks. The chlorite associated with biotite is isotropic or faintly blue under crossed nicols, and the index of refraction of cleavage flakes is higher than that for chlorite with hornblende. The optical properties correspond most closely to aphrosiderite (Winchell, 1933). The chlorite often shows a radial or spherulitic structure and includes magnetite. This variety is widely distributed but not abundant.

Much of the chlorite associated with hornblende is a faint dirty brown under crossed nicols, and the indices are lower than those for chlorite associated with biotite. The optical properties suggest the variety ripidolite (Winchell, 1933).

SPHENE: Sphene is widespread as an accessory. The hornblende-bearing and andesine-bearing rocks contain more sphene than the other rocks of the area. The dark-brown to gray-brown crystals are euhedral to subhedral and are 3 millimeters in diameter in some hornblende gneisses (Fig. 10). Small dark grains lie between the contacts of hornblende, biotite, and epidote. Some sphene crystals are granulated.

PYRITE: This is also a common accessory. The hornblende gneisses contain more pyrite than the other rock types. Pyrite grains are irregular or euhedral, and they contrast with the anhedral pyrite of the schists. Some pyrite is intergrown with epidote. Pyrite alters to limonite and hematite, principally by weathering.

APATITE: Apatite is a common accessory. The colorless crystals are slightly larger than most sphene crystals. Many of their corners are rounded or broken, and the fragments are slightly displaced. The apatite in the pegmatites is granulated. Green apatite from a cassiterite-bearing pegmatite near Rockford is negative, with $\omega = 1.642$ and $\epsilon = 1.637$.

OTHER ACCESSORIES: Zircon is rare except as inclusions in biotite and hornblende, and it is always surrounded by pleochroic haloes.

Garnet occurs sporadically in very minor amounts and is usually fractured. Some garnets are subhedral. In epidote-quartz-garnet rocks, the garnet is poikiloblastic toward quartz and epidote.

Tourmaline is not common but occurs in pegmatites and aplites. Large crystals occur in characteristic long prisms in the pegmatites, and in fine-grained rocks they are small and stumpy. The tourmaline is black, and the indices indicate schorlite. Tourmaline increases slightly toward the margins of the Pinckneyville complex.

Traces of carbonate grains are scattered throughout all rocks. The carbonate is colorless and interstitial. Twinning and cleavage planes are not visible.

Sericite is infrequently found in plagioclase as matted, fibrous patches which do not occupy the entire grain.

The opaque minerals include hematite, limonite, and leucoxene as weathering products of pyrite and sphene. Magnetite develops in the alteration of biotite.

Except for pyrite and magnetite resulting from the alteration of biotite, the opaque ore minerals are rare.

Sequence of crystallization in the Pinckneyville complex.—There is no well-defined paragenetic mineral sequence for the igneous rocks of the Pinckneyville complex, and the crystallization of many of the minerals seems to have been sudden and overlapping and influenced by the intrusion of the magma under conditions of regional metamorphism. Some data, however, do indicate a general sequence of crystallization.

That the plagioclase is primary is shown by the normal zoning and twinning, the large subhedral outlines, and the protoclastic deformation of the crystals. Plagioclase crystallized early as is shown by the phenocrystlike habit, the arrangement of the other minerals with respect to plagioclase rather than plagioclase to them, and the stronger protoclastic deformation in the rocks with more pronounced flow structures.

Zoning is better developed in the oligoclase of the granodiorite than in the andesine of the quartz diorite. Zoning indicates a somewhat rapid growth and failure to attain complete equilibrium. Intrusion would tend to lessen the opportunity for complete reaction of the crystals in the magma. The quartz diorite is the earliest member of the complex, and its early crystallization probably took place under more quiescent conditions than that of the granodiorite which followed it.

Epidote, biotite, and the accessories probably started to crystallize slightly later than plagioclase. Epidote seems to be earlier, in part at least, than biotite. The including of the accessories by epidote and biotite aggregates indicates that they developed about the same time. Some muscovite appears to have been formed with biotite. Muscovite in general probably had a longer crystallization history than biotite since its ratio to biotite increases in the more acid rocks, and it is one of the principal constituents of the later pegmatites.

Potash feldspar is late because it replaces plagioclase and is interstitial to all minerals except quartz. Quartz may also have crystallized for a long period of time but it was the last mineral to cease crystallization.

Origin of epidote.—Epidote is a characteristic and a widespread mineral in the rocks of the Pinckneyville complex. It is not among the usual primary minerals of the granitic types of igneous rocks, although Johannsen (1934) describes several quartz diorites and granodiorites containing amounts of epidote comparable to those in the Pinckneyville complex.

Several writers believe that some epidote in igneous rocks is primary. Krauskopf (1941) considered some of the epidote in the Osoyoos batholith of Washington to be

primary, partly because it was earlier than biotite. Osborne (1934) ascribed a probable primary origin for epidote in granitic rocks in Quebec and considered it a stress mineral, perhaps developed in the rock as a result of deformation. Earlier workers also considered some epidote as primary (Butler, 1909; Lacroix, 1889; and others).

Eskola (1914) described some granitic rocks in Finland which contain epidote, allanite, and biotite and whose relationships are identical with those in the Pinckneyville complex. He discarded the hypothesis that epidote and biotite resulted from the breakdown of potash feldspar and the interaction of its products with some (MgO, FeO)- and (CaO)-bearing mineral such as an amphibole or pyroxene. He argued that (1) the alumina content of biotite plus epidote is greater than that of potash feldspar plus any mafic mineral available, and (2) the ratio of biotite to epidote was variable. Both arguments apply to the Pinckneyville, and furthermore there are no relict structures which could suggest such a reaction. Eskola believed that both minerals were of metamorphic origin and that the allanite grains were centers of crystallization. He suggested that the only other possibility for formation of epidote was through the breakdown of anorthite.

Epidote (1), in part at least, is earlier than the biotite, and its euhedral outline suggests early growth. The allanite always occurs as cores in epidote (1) and seems to have served as centers of growth for much of epidote (1). Allanite and epidote are members of the same mineral family, and allanite is generally considered a primary accessory of igneous rocks.

It is, therefore, suggested that epidote (1) is primary and due in part to crystallization in a magma intruded during folding under the stress of regional metamorphism. Epidote (1) may have crystallized instead of some plagioclase feldspar, and epidote (3) developed as an alteration product of plagioclase at the same time as epidote (1) because epidote was a stable phase at that time. The many epidote inclusions in the interior of plagioclase crystals seems to support this because it intimates an early alteration of plagioclase prior to complete crystallization of the feldspar.

Rocks.—Quartz diorite is the principal rock type of the Pinckneyville complex, with minor granodiorite, granite, and hornblende gneiss. Several types of hornblende gneiss are exposed and are thought to represent metamorphosed basic rocks intruded earlier than the Pinckneyville.

The sequence of intrusion of the Pinckneyville is quartz diorite, granodiorite, and granite. Feldspar, quartz, biotite, and muscovite are the essential minerals. The normal rock types are dark gray to white and coarse-grained. Rock structures and textures are similar for all normal types and, the textures grade from subporphyritic to strongly gneissic.

Dikes, pegmatites, and quartz veins cut the complex. Inclusions are abundant in some rocks. Other rock types in the complex are injection gneisses, mylonites, and altered hornblende gneiss.

HORNBLENDE GNEISS: The hornblende-bearing rocks are included here although their origin is uncertain. This group is divided into three types: (1) amphibolite, (2) hornblende-epidote gneiss, and (3) hornblende-biotite gneiss. Hornblende, plagioclase, and epidote occur in all types in varying proportions. Chlorite occurs in rocks with little or no biotite, and most of it is an alteration product of hornblende.

(1) The amphibolites are dark green and consist of about 70 per cent hornblende

and varying proportions of feldspar, epidote, and chlorite. The hornblende crystals are rectangular and blocky, 1 centimeter or more in diameter. A typical mode is given in Table 1 (No. 6). Plagioclase ranges from calcic albite to sodic oligoclase with traces of potash feldspar. Quartz is a minor constituent. Sphene is the most common accessory, and others are carbonate, apatite, biotite, and opaque black minerals.

The rocks are massive but show a very faint megascopic and microscopic orientation of hornblende. Plagioclase fills the interstices and outlines the large hornblende crystals. In thin section small hornblende grains are intimately intergrown with epidote and chlorite but are idiomorphic toward all other minerals. Chlorite is concentrated between large hornblende grains or occurs as thin light-green borders on hornblende (Fig. 10) and includes magnetite. Epidote is subhedral where associated with hornblende and makes euhedral inclusions in plagioclase. Strings of sphene crystals follow grain boundaries or cleavage traces.

(2) Modes of hornblende-epidote gneisses are given in Table 1, Nos. 7 and 8. This type is light green, particularly where feldspar is abundant, but grades into the darker gray-green hornblende-biotite gneisses. The essential minerals are hornblende, plagioclase, epidote, chlorite, and quartz with accessory sphene and pyrite. The plagioclase is usually oligoclase or andesine, and there are traces of potash feldspar. The plagioclase includes epidote crystals and a few small mica laths.

Hornblende and epidote are subhedral with slightly better crystal outline than feldspar. Chlorite is pseudomorphic after hornblende or is concentrated in the interstices. It includes some magnetite. Quartz is anhedral and interstitial. A few ragged grains of garnet enclose quartz and epidote.

The rocks are medium- to fine-grained and as a rule show a lineation. Planar structures occur locally. All minerals participate in the structures. In a few thin sections small granulated and sheared zones were observed.

(3) The composition and texture of the hornblende-biotite gneisses are variable, and the rocks grade into the hornblende-epidote gneisses. The minerals are hornblende, plagioclase, biotite, quartz, and epidote. Accessories are sphene, apatite, pyrite, allanite, and zircon. Interstitial carbonate also occurs. Modes of typical material are given in Table 1, Nos. 9–12. Figure 11 is a sketch of No. 11.

Plagioclase ranges from andesine to labradorite. The more basic plagioclase occurs where biotite is an essential constituent. The grains are large and blocky or small and irregular. Inclusions in plagioclase are white mica, epidote, biotite, and hornblende. The first two are common and follow crystallographic directions. Rounded or irregular patches of matted sericite occur in some plagioclase grains. The feldspars are fractured, and their corners are broken.

Hornblende is always subhedral and is closely associated with biotite and epidote. In some instances the latter two minerals are included poikiloblastically. Hornblende is larger than all minerals except blocky plagioclase. Epidote is scattered throughout the rocks in which it is abundant, but generally it is associated with biotite. The biotite is brown, green brown, or red brown.

The structure is gneissic, and the larger feldspars occur in augen. There are all gradations between well-developed augen in well-foliated rocks and poorly developed

augen in rocks with a good lineation. This structure is emphasized by inequidimensional biotite patches with their longest dimension parallel to the lineation. Individual biotite grains within these patches are subparallel.

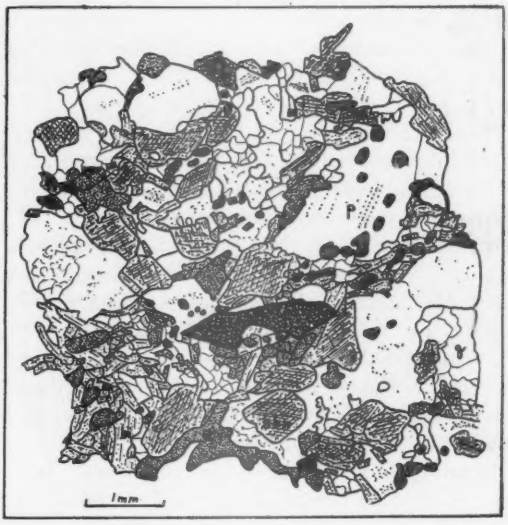


FIGURE 11.—*Hornblende-biotite gneiss*

With large euhedral sphene crystal; sec. 24, T. 20 N., R. 18 E., Elmore County. (hbn) hornblende, (bi) biotite, (p) plagioclase, (q) quartz, (e) epidote

ORIGIN OF HORNBLENDE GNEISSES: There are four possible explanations for the origin of the hornblende gneisses. They may have been derived from: (1) original sedimentary beds, perhaps impure carbonates, (2) basic intrusions belonging to the Pinckneyville complex, (3) basic lava flows, or (4) basic sills and dikes of an earlier intrusion. Prouty (1923a) suggested both sedimentary and igneous origins for similar rocks in Clay County, but he did not distinguish clearly which of the two was applicable. Brown (1925) discussed the possibilities of sedimentary and igneous origins for hornblende gneisses from Clay County, and he concluded that they probably represented basic sills and dikes.

Hornblende gneisses and schists are widely distributed in Alabama and occur in small lenses to beds of considerable width in schist from Alabama into Georgia (C. F. Park, Jr., personal communication) and probably continue to the northeast. The occurrence of pyroxene and olivine and the association of cortlandite with certain hornblende rocks in Clay County (Brown, 1925) suggests that they are, in part at least, of igneous origin.

Although the geographic distribution seems to support the sedimentary origin, this would necessitate the complete alteration of impure carbonate rocks to hornblende gneisses and schists, because no relicts of carbonate rock have been found in this area, and typical metamorphic minerals such as tremolite, wollastonite, diopside,

and others are present in very minor amounts or are lacking. The schist series of this area is composed of clastic metasediments in which carbonate beds do not occur in significant amounts. The sedimentary explanation also fails to explain the cortlandite and olivine.

Quartz diorite includes hornblende gneiss, and the well-developed structure and the degree of metamorphism of the inclusions indicate they are earlier than the quartz diorite. The geographic distribution of hornblende rocks precludes the possibility that they are part of the Pinckneyville complex.

The intercalation of hornblende rocks with schist is more readily explained by dike or sill intrusion than by basic lava extrusion. In view of the available data, it seems most likely that these hornblende gneisses represent metamorphosed basic sills and dikes intruded earlier than the Pinckneyville intrusion.

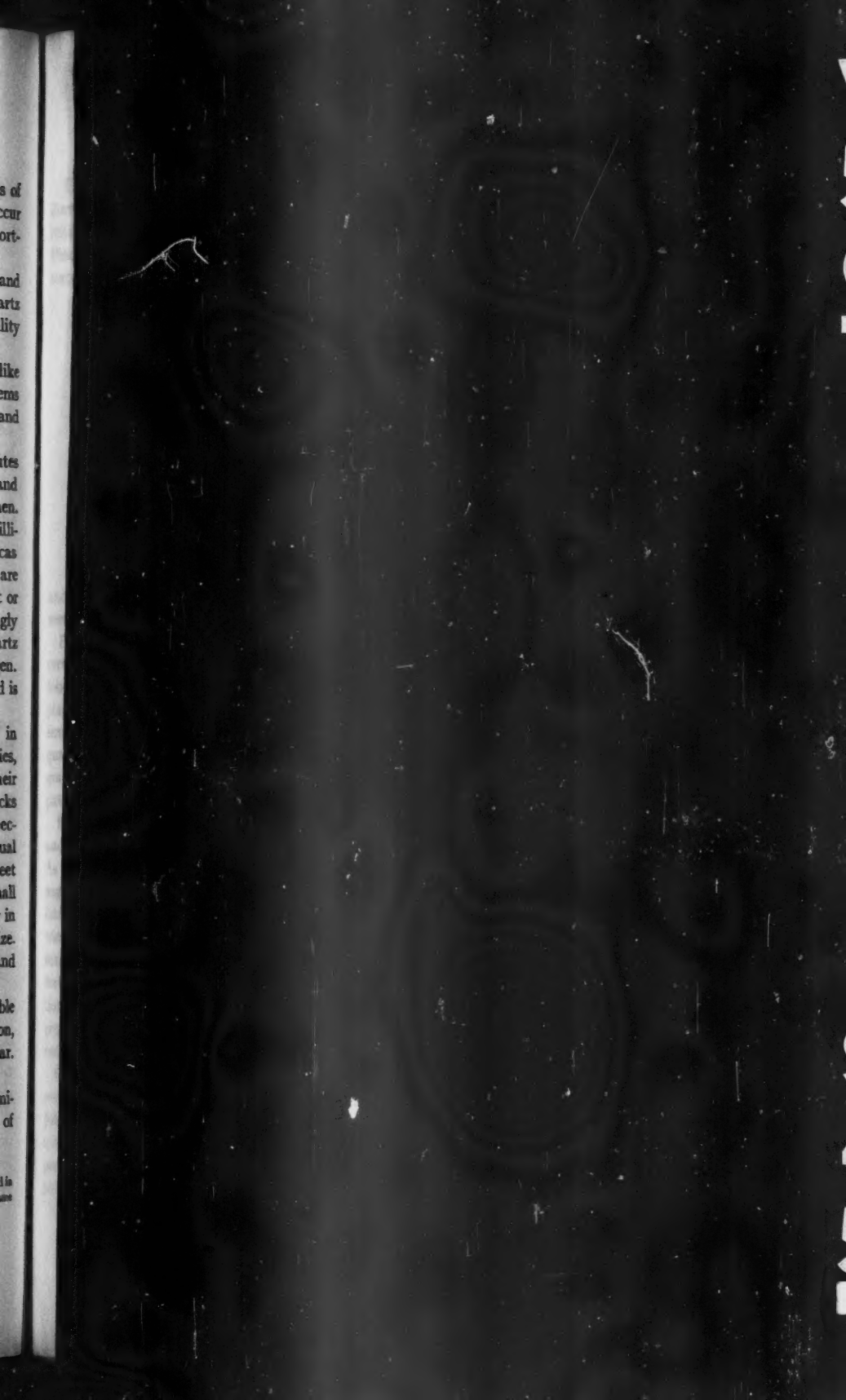
QUARTZ DIORITE GNEISS:[2] This is the typical Pinckneyville rock and constitutes most of the complex. The normal rock is dark to medium gray, coarse-grained, and contains plagioclase, biotite, muscovite, and quartz recognizable in hand specimen. The plagioclase grains are white and blocky. Their diameter averages 3 to 4 millimeters or about 5 to 10 times that of other minerals. They are surrounded by micas and quartz, and where lineation or foliation is well developed the feldspar grains are in distinct augen. A lineation is usually visible, but foliation may be indistinct or lacking. The poorer the foliation, the less pronounced the augen. In strongly foliated rocks, the feldspars are strung out in long lenses. In such cases quartz makes thin tabular lenses which bend and weave with the micas to outline the augen. In weathered outcrops this quartz stands out in relief, and the rock is bluish and is known locally as "blue granite."

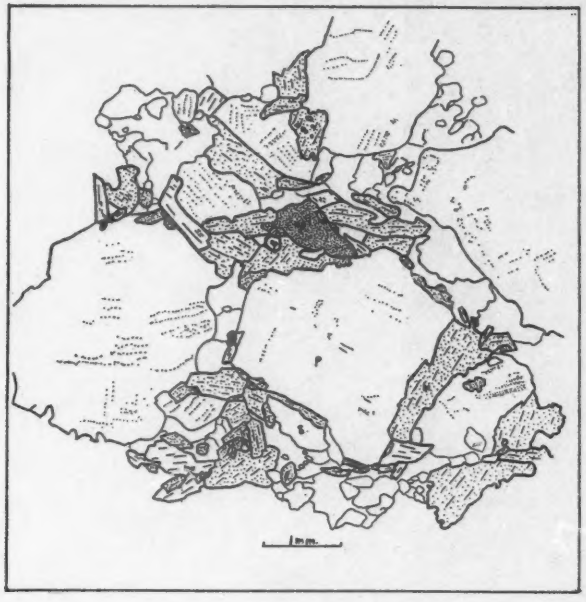
There are many varieties of quartz diorite gneiss, and they differ principally in texture and structure. The minerals are uniformly distributed in most varieties, particularly in the more massive rocks. Fine-grained varieties are darker, and their lineation is more prominent than their foliation. In some flat outcrops the rocks are uniform and massive except for the lineation. In other exposures, close inspection is necessary to distinguish the slightly contrasting bands resulting from gradual or sharp textural changes in grain size. The bands are several inches to several feet or more wide. If the variations are sharp, some of the structures crosscut at small angles (Fig. 16). Irregular, strongly contrasting light and dark bands also occur in some outcrops due to varying amounts of biotite and slight changes in grain size. These bands are 2 inches to 1 foot wide and grade into one another parallel to and across the strike (Fig. 12).

The essential minerals are plagioclase, quartz, and biotite with smaller, variable quantities of muscovite and epidote. Accessory minerals are apatite, sphene, zircon, pyrite, allanite, and occasionally garnet, carbonate, and traces of potash feldspar. Modes of typical material are given in Table 1.

Plagioclase ranges between Ab_{55} and Ab_{75} with andesine (about Ab_{60-65}) predominating. Potash feldspar increases with increasing sodic plagioclase. Patches of quartz grains may reach the size of plagioclase grains.

[2] The quartz diorites may also be called tonalites (Johannsen, 1934; Harker, 1935). The term quartz diorite is used in this report because previous workers in this general region have applied the name quartz diorite to rocks of the same composition which are probably offshoots of the main Pinckneyville complex.



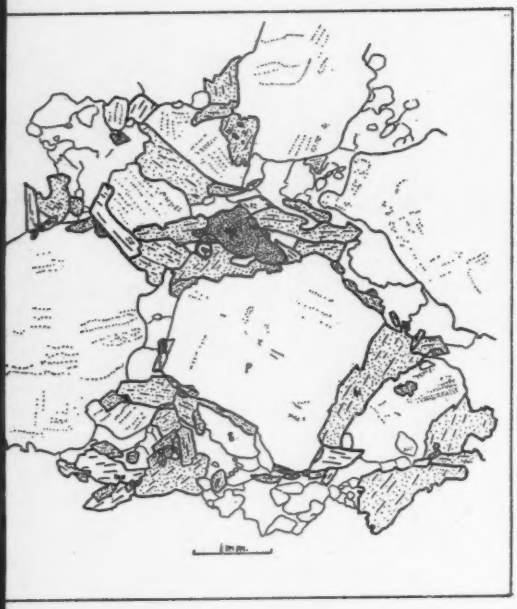


a

a. Quartz diorite; sec. 20, T. 24 N., R. 21 E., Tallapoosa County.
b. Quartz diorite; sec. 1, T. 23 N., R. 20 E., Coosa County.
c. Quartz diorite; sec. 9, T. 20 N., R. 17 E., Elmore County (D 9–11).
d. Quartz diorite; sec. 18, T. 22 N., R. 18 E., Coosa County (D 30–32).
e. Quartz diorite; sec. 16, T. 20 N., R. 18 E., Elmore County (D 4–6).
f. Quartz diorite; sec. 11, T. 20 N., R. 18 E., Elmore County (D 27–29).
g. Quartz diorite; sec. 21, Elkahatchee Creek, Tallapoosa County.
h. Granodiorite border gneiss, Elkahatchee Creek, Tallapoosa County (D 56–58).

e

Camera lu



a

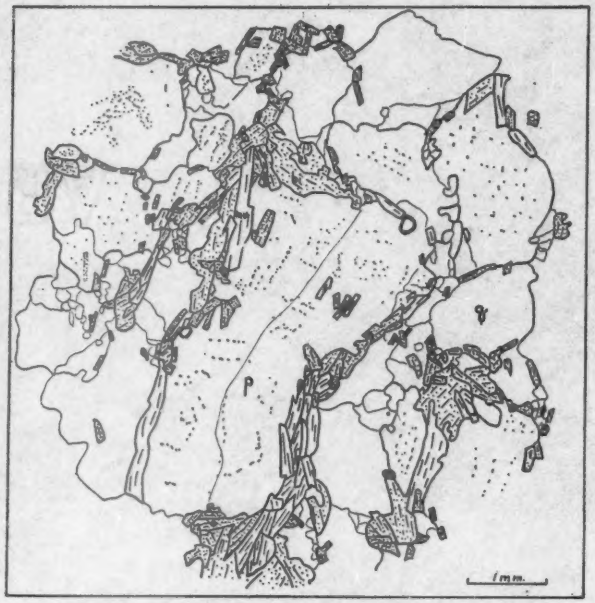
b

e

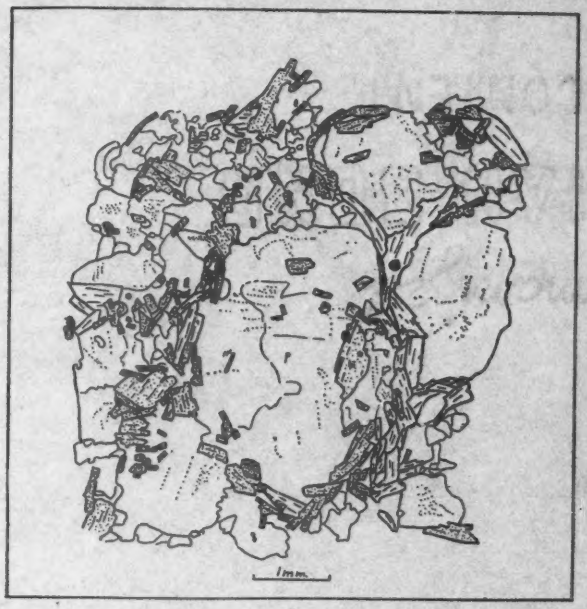
f

TEXT

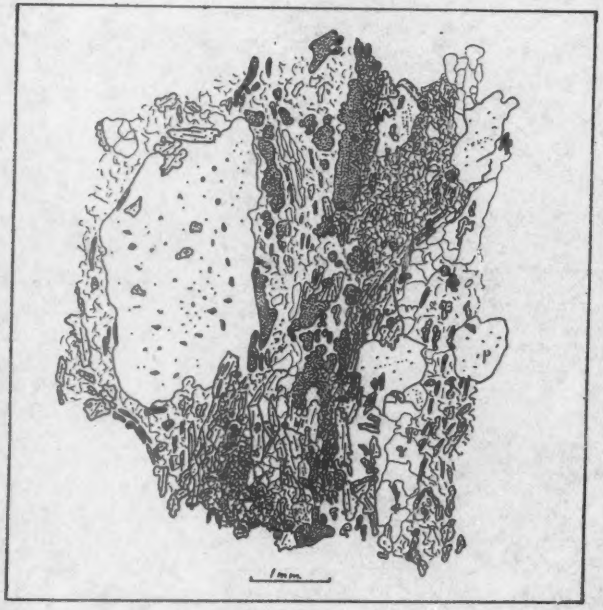
Camera lucida sketches of types of textu



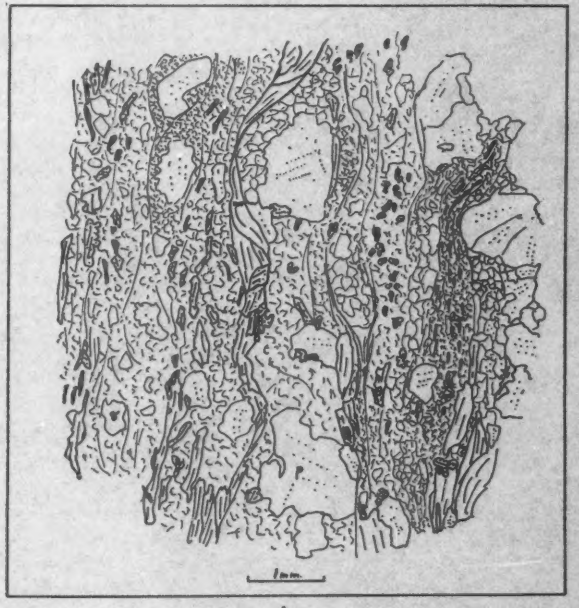
c



d



g



h

TEXTURAL TYPES IN PINCKNEYVILLE COMPLEX

s of textures in quartz diorites and granodiorite from slightly granulated to intensely granulated

The texture is inequigranular and subporphyritic. Plate 2 shows a series of sketches grading from massive and poorly foliated material (Pl. 2, a) to strongly foliated material (Pl. 2, h). The textures illustrated are characteristic of the normal Pinckneyville quartz diorite gneiss, and the general features are representative of the textures in other rock types of the complex. Differences between these textures

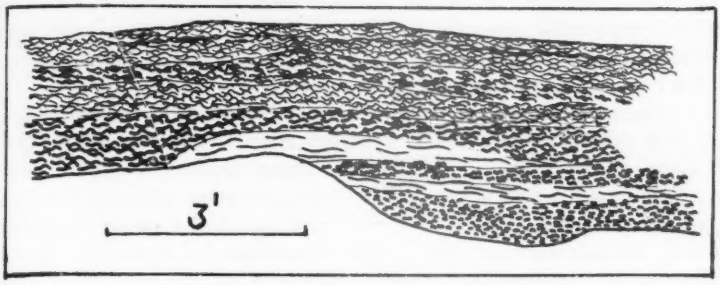


FIGURE 12.—*Banding in quartz diorite gneiss*
Bands are not continuous; sec. 30, T. 20 N., R. 18 E., Elmore County.

and those in other rock types, for example dike rocks, are in the degree of the development of the foliation and the grain size.

Plate 2 illustrates textural features more clearly than they can be described, but certain points are worthy of mention. The grain size of all minerals decreases as foliation becomes more intense. This is more apparent in biotite and quartz than in plagioclase in the early stages. The subporphyritic texture grades into an augen texture which is more prominent as the foliation becomes more intense. Mica and quartz patches are more elongate in the finer-grained textures. The individual mica grains are not always oriented parallel to the long direction of the patches, and the patches are not always parallel to the foliation.

Feldspar crystals gradually lose their subhedral shape by granulation of the corners and edges, and the undeformed parts of the crystals become the centers of the augen. As the augen grow more distinct and lens-shaped, the feldspar crystals within the augen are fractured, the fragments rotated, and each fragment may be broken further into smaller grains. The zone of granulated material surrounding the feldspar widens, but the size of the augen does not change markedly. The strong gneissic texture of Plate 2, h is uncommon and represents a highly granulated phase only 20 feet from the main eastern schist contact. It illustrates the intensity of the deformation near contacts at some localities where resistance to flow of the crystals was greatest. Plate 2, g shows the texture of material only 880 feet away from the same contact.

GRANDIORITE GNEISS: The terms granodiorite and granite as used here include several rock types of the granodiorite (family 7) and granite (family 6) families of Johannsen (1934). Since exhaustive petrographic and petrologic studies to determine all possible rock types were not made, no further separation of rock types seems permissible. The division between granodiorite and granite is made where the potash feldspar content constitutes more or less than 50 per cent of the total feldspar.

Granodiorite gneisses are widely distributed but less abundant than quartz diorite.

They are light gray to buff and white and are more leucocratic than the quartz diorites. The darkness of the rock is an index to the amount of biotite. Feldspars, quartz, and mica are visible in hand specimen, but it is difficult to identify the feldspar. The grain size is about the same as in the quartz diorites, but fine-grained varieties are more abundant.

Plagioclase, microcline, orthoclase, quartz, muscovite, and biotite are the chief constituents with small amounts of epidote. Accessories include allanite, apatite, and zircon with subordinate sphene and pyrite. Occasional garnet crystals occur. Modes of typical material are given in Table 1.

The plagioclase is usually oligoclase but ranges between sodic andesine and calcic albite. The size, shape, and internal structures of the feldspars are similar to those in the quartz diorites. Zoning is slightly more pronounced, but twinning is less distinct. Grains with well-developed albite twins are more subhedral than poorly twinned grains. Secondary growth of calcic albite on plagioclase grains occurs here and there. Inclusions in plagioclase are epidote and mica, but mica is more prevalent.

Microcline and orthoclase occur predominantly in medium-sized grains with anhedral outline but also as fine granular material mixed with quartz or small plagioclase grains. The granular areas are in the large spaces between plagioclase phenocrysts and are abundant in a few rocks. Potash feldspars replace plagioclase, but the replacement is usually not complete. Myrmekite is abundant in some rocks and almost lacking in others. The usual myrmekite is small wormy quartz in plagioclase, but occasionally the intergrowth is graphic. The quartz blebs are in optical continuity.

Biotite and epidote are less abundant than in the quartz diorite gneiss. Muscovite is more abundant, and the amount of quartz is about the same. Epidote includes allanite cores. Muscovite (3) and (4) are common in the granodiorites and replace plagioclase. The replacement textures are myrmekitic or porphyroblastic (Figs. 7b, 8a). The fibers cut across grain boundaries or grow on other muscovite grains (muscovite 1) and are not, apparently, related to the rock structure.

Apatite, biotite, muscovite, epidote, and plagioclase are fractured or partially granulated.

Granodiorite textures and structures are similar to those of the quartz diorite gneiss, but taken as a group they are not so pronounced except in certain localities. Although Plate 2 illustrates quartz diorite textures, comparable sketches could be made of granodiorite textures. The textural relations are affected only slightly by the different proportions of feldspars, biotite, muscovite, and epidote in the granodiorites.

GRANITES: All the granites contain considerable plagioclase, but potash feldspar always constitutes more than 50 per cent of the total feldspar. Ganites are not abundant, and they are difficult to distinguish from the granodiorites into which they grade imperceptibly. Many of the leucocratic rocks are granites.

The granites consist chiefly of microcline, orthoclase, sodic oligoclase or calcic albite, quartz, and muscovite with minor amounts of biotite and epidote. Accessories are apatite, tourmaline, garnet, allanite, pyrite, and sphene. The decreasing ratio of dark minerals to potash feldspar from quartz diorite to granodiorite continues in the granites. Modes of two granites are given in Table 1.

TABLE 1.—*Modes of some metamorphic and ign*

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	16a
Plagioclase	15.2		18.1		3.9	9.6	31.0	57.8	34.9	43.9	33.5	33.6	44.8	47.3	48.4	38.9	35.7
Potash Feldspar														1.5			
Quartz	46.9	9.4	18.9	17.8	54.4	1.0	13.0	4.8	6.0	20.6	16.7	22.3	25.2	29.3	31.3	29.3	16.2
Hornblende						71.9	22.9	14.2	45.5	11.3	19.2	11.4					
Biotite	29.6	.7	47.6	47.8	20.2		tr.		9.5	16.6	17.2	21.1	17.3	12.9	11.8	20.4	21.5
Epidote	3.6	4.0	4.1			12.6	21.5	6.9	2.3	5.6	11.0	2.6* 6.9**	7.3	1.9	6.4	2.2	12.1
Muscovite	2.3	38.2	11.3	13.5	19.2								3.1	5.5	1.9	8.3	12.8
Chlorite		35.0				4.2	6.8	6.7									
Sillimanite				20.9													
Accessories‡	2.3	tr.	tr.	tr.	2.3		4.8		.6						.2	.9	tr.
Sphene						.7		3.1	1.2	1.3	1.7	1.6	.5	1.3			
Pyrite		10.5								.7			.9				
Vein material								6.5									
Apatite											.6		.5				
Carbonate												.5	.4	.3			
Garnet		2.2															
Totals.....	99.9	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	99.9	100.0	100.0	100.0	100.0	100.0	100.0

* in groundmass.
** included in feldspar.
† phenocrysts.
‡ includes all accessories unless otherwise shown.

1. Fine-grained, dark-gray, laminated rock suggesting quartzite in hand specimen; sec. 10, T. 22 S., R. 7 E., Clay Co.
2. Dark, coarse-grained crumpled schist with muscovite porphyroblasts; sec. 1, T. 24 N., R. 21 E., Tallapoosa Co.
3. Fine-grained dark schist inclusion; Elkahatchee Creek, Tallapoosa Co.
4. Coarse-grained, dark, crumpled sillimanite schist; sec. 27, T. 24 N., R. 21 E., Tallapoosa Co.
5. Coarse-grained, dark quartz biotite schist; sec. 17, T. 23 N., R. 20 E., Coosa Co.
6. Amphibolite; sec. 8, T. 20 N., R. 19 E., Elmore Co.
7. Hornblende-epidote gneiss, sec. 8, T. 22 S., R. 7 E., Clay Co.
8. Hornblende-epidote gneiss, sec. 36, T. 22 N., R. 20 E., Coosa Co.
9. Hornblende-biotite gneiss; sec. 6, T. 21 N., R. 19 E., Coosa Co.
10. Hornblende-biotite gneiss; sec. 8, T. 23 N., R. 20 E., Coosa Co.
11. Hornblende-biotite gneiss; sec. 24, T. 20 N., R. 18 E., Elmore Co.
12. Hornblende-biotite gneiss; Elkahatchee Creek, Tallapoosa Co.
13. Quartz diorite gneiss; sec. 31, T. 23 N., R. 20 E., Coosa Co.
14. Quartz diorite gneiss; sec. 1, T. 23 N., R. 20 E., Coosa Co.
15. Quartz diorite gneiss; near Martin Lake; sec. 2, T. 21 N., R. 20 E., Coosa Co
16. Quartz diorite gneiss; sec. 10, T. 23 E., R. 21 E., Tallapoosa Co.

16a. Inclusion in quartz diorite gneiss; sec. 10, T. 23 N., R. 21 E., Tallapoosa Co.

17. Quartz diorite gneiss; Elkahatchee Creek, Tallapoosa Co.

17a. Fine-grained, dark band adjacent to 17; Elkahatchee Creek, Tallapoosa Co.

ABLE 1.—*Modes of some metamorphic and igneous rocks of the Pinckneyvi*

	Locality														
9	10	11	12	13	14	15	16	16a	17	17a	18	19	20	21
4.9	43.9	33.5	33.6	44.8	47.5 1.5	48.4	38.9	35.7	48.6	30.4	42.8	46.2 .5	37.4	66.
6.0	20.6	16.7	22.3	25.2	29.3	31.3	29.3	16.2	31.4	29.7	28.1	23.7	21.0	25.
5.5	11.3	19.2	11.4											
9.5	16.6	17.2	21.1	17.3	12.9	11.8	20.4	21.5	12.1	26.5	6.9	18.6	29.3	4.
2.3	5.6	11.0	2.6* 6.9**	7.3	1.9	6.4	2.2	12.1	8.9	13.4	5.6	2.5	3.4	1.
				3.1	5.5	1.9	8.3	12.8			16.6	5.9	7.0	1.
.6						.2	.9	tr.	tr.	tr.	tr.		.3	.
.2	1.3	1.7	1.6	.5	1.3									
	.7			.9										
		.6		.5								.1		
			.5	.4	.3							1.9		
.0	100.0	99.9	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.

ende-biotite gneiss; sec. 8, T. 23 N., R. 20 E., Co.

ende-biotite gneiss; sec. 24, T. 20 N., R. 18 E., Co.

ende-biotite gneiss; Elkahatchee Creek, Tallapoosa

diorite gneiss; sec. 31, T. 23 N., R. 20 E., Coosa Co.

diorite gneiss; sec. 1, T. 23 N., R. 20 E., Coosa Co.

diorite gneiss; near Martin Lake; sec. 2, T. 21 N., ., Coosa Co

diorite gneiss; sec. 10, T. 23 E., R. 21 E., Tallapoosa

ion in quartz diorite gneiss; sec. 10, T. 23 N., R. allapoosa Co.

diorite gneiss; Elkahatchee Creek, Tallapoosa Co.

rained, dark band adjacent to 17; Elkahatchee Tallapoosa Co.

18. Quartz diorite gneiss; sec. 14 Co.
19. Quartz diorite gneiss; sec. 5,
20. Quartz diorite gneiss; Speig R. 18 E., Elmore Co.
21. Quartz diorite gneiss; Socap
22. Granodiorite gneiss, (leuc 10, T. 22 N., R. 21 E., Talla
23. Granodiorite gneiss, (adame Tallapoosa Co.
24. Granodiorite gneiss, (mon R. 21 E., Tallapoosa Co.
25. Granodiorite gneiss, (ada Tallapoosa Co.
26. Granodiorite gneiss, (monz ford, Coosa Co.
27. Granodiorite porphyry (Creek, Tallapoosa Co.

Pinckneyville complex, Alabama

																Average Composition		
	20	21	22	23	24	25	26	27	28	29	30	31	32	33	35	36	37	38
:2	37.4	66.8	38.7	30.8	48.4	23.1	41.7	28.6* 13.8†	38.9	23.5	20.6	16.6	32.6	44.5	36.4	34.9	45.2	37.7
.5			8.2	22.5	10.4	21.7	12.0	7.1	13.4	36.3	26.2	29.5	26.8		3.8		.2	13.6
.7	21.0	25.8	35.8	30.8	28.2	26.5	27.1	25.4	31.2	25.4	31.6	35.5	31.6	21.0	41.1	12.1	27.5	29.3
																28.1		
6	29.3	4.5	1.6	6.8	2.3	11.1	7.8	7.2	11.5	1.1	4.8	5.3	1.3	30.8	7.4	9.2	16.0	6.9
.5	3.4	1.0	2.8	3.3	5.4	4.4	2.7* .5**	1.0	2.2	1.2	1.7	2.2		2.2		9.9	5.3	3.2
9	7.0	1.4	12.0	5.8	5.3	12.7	6.4* 1.8**	16.7	2.8	12.5	15.1	10.8	7.7	1.5	11.3		5.0	9.1
																2.5		
	.3	.5	tr.	tr.	tr.	.5	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	1.7	.8	.2
			.5													1.6		
1																		
9																		
			.5															
0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	99.8	100.0

gneiss; sec. 14, T. 24 N., R. 21 E., Tallapoosa

gneiss; sec. 5, T. 19 N., R. 18 E., Elmore Co.
gneiss; Speigner's quarry; sec. 11, T. 20 N., re Co.
gneiss; Socapatoy Creek.
gneiss, (leucomonzatonalite), quarry; sec. 21 E., Tallapoosa Co.
neiss, (adamellite); sec. 2, T. 23 N., R. 21 E.,

gneiss, (monzatonalite); sec. 1, T. 23 N., apoosa Co.
gneiss, (adamellite); Elkahatchee Creek,

neiss, (monzatonalite); D 41 – D 44, Rock-
porphyry (monzatonalite); Elkahatchee osa Co.

28. Granodiorite gneiss (monzatonalite); sec. 30, T. 20 N., R. 18 E., Elmore Co.
29. Granite gneiss (leucoadamellite); sec. 10, T. 23 N., R. 20 E., Coosa Co.
30. Granite gneiss (adamellite); sec. 17, T. 22 N., R. 20 E., Coosa Co.
31. Dike rock (monzogranite); sec. 14, T. 20 E., R. 17 E., Elmore Co.
32. Dike rock (leucoadamellite); sec. 5, T. 20 N., R. 19 E., Elmore Co.
33. Quartz diorite inclusion; sec. 14, T. 20 N., R. 17 E., Elmore Co.
35. Fine-grained crumpled injection gneiss; sec. 24, T. 20 N., R. 18 E., Elmore Co.
36. Average of 6 to 12 inclusive.
37. Average of 13 to 21 inclusive.
38. Average of 22 to 28 inclusive.

Rock textures and structures are similar to those of the granodiorite gneisses.

DIKE ROCKS: Dikes are abundant. They are light gray, buff and white, and the color depends upon the amount of biotite and the grain size. The rocks are medium- to fine-grained and finer-grained than the granite, granodiorite, and quartz diorite. Quartz, feldspar, and mica are recognizable in hand specimen, and occasionally small garnet or tourmaline crystals. A few dikes contain thin biotite aggregates, which are probably included material derived from the dike walls but are not always oriented parallel to dike walls or foliation.

The chief minerals are plagioclase, orthoclase, microcline, quartz, and muscovite with minor amounts of biotite and epidote. Accessory minerals are allanite, garnet, tourmaline (rare), apatite, zircon, and pyrite. The plagioclase is predominantly oligoclase but ranges from sodic andesine to calcic albite. Carlsbad twins are abundant. All dikes are granodiorites or granites. The aplites do not carry biotite. Modes of two dikes are given in Table 1.

The dike rocks are (1) equigranular or (2) porphyritic. The equigranular textures are similar to those of the granodiorites. Planar structures are present in all dike rocks but are less well developed in the rocks with much potash feldspar. In the porphyries, the phenocrysts are large feldspar crystals in augen, and many crystals are fractured internally. A few augen are composed of granulated phenocrysts. The phenocrysts are outlined by muscovite which is either in direct contact with the crystal or is separated by a narrow zone of fine-grained quartz and feldspar.

Contacts with the wall rock are sharp but slightly irregular because the surfaces of the fractures in the inequigranular wall rock are rough.

PEGMATITES AND QUARTZ VEINS: Pegmatites in the gneissic rocks are very numerous. The widths of pegmatites is variable, rarely exceeding 20 inches and as narrow as small fractions of an inch. Grain sizes vary from less than a millimeter to more than 10 centimeters. The pegmatites are light pink, gray, and white.

The chief minerals are oligoclase, albite, orthoclase, microcline, quartz, and muscovite. Accessory minerals are allanite, garnet, tourmaline, biotite, and in addition apatite, beryl, and cassiterite in the pegmatites in the schist belt on the western side of the complex. Graphic intergrowth of quartz and microcline is rare. Some pegmatites have quartz centers and quartz and feldspar borders.

Foliation is poorly developed in the pegmatites, partly as a result of the coarse grain of the rocks. In some the quartz and feldspar grains are elongate or lenslike and are outlined by thin books of muscovite. Many grains are granulated, and the muscovite flakes or books are bent and sheared. Small garnets are fractured and occasionally appear granulated and rolled out into strings of grains.

Quartz veins are very common but small. In the schists quartz veins are 10 feet or more thick and parallel the schistosity. In gneissic rocks they are 0.1–1 inch wide and several feet long. They parallel the strike of the foliation or lineation or cut across at any angle. Infrequently quartz segregates in patches 4 to 8 inches in diameter.

INCLUSIONS: There are many inclusions in some rocks, but none in others. The quartz diorite contains more inclusions than any other rock type. The most common type of inclusion is a fine-grained, dark, gray-blue rock very similar to the darkest fine-grained varieties of quartz diorite. The inclusions range from an inch wide and

6 inches long to an inch wide and 12 feet long in flat outcrops, and appear to be disc- or oval-shaped.

Plagioclase, biotite, and quartz are the main constituents with smaller amounts of epidote and muscovite. Allanite, pyrite, sphene, and apatite are accessories. Modes are given in Table 1, Nos. 33 and 16a.

The plagioclase is a calcic andesine or sodic labradorite. Some muscovite occurs as porphyroblasts. The textures and mineral relations are similar to those in the quartz diorite gneiss, but the micas are more bent and folded (Fig. 7b). The finer grain size, darker color, and larger amounts of biotite and epidote distinguish them from quartz diorite. The edges of the inclusions are rounded. There is no contact zone developed either in the inclusion or in the enclosing rock, but the surfaces of many inclusions are coated with biotite.

Other inclusions are hornblende gneiss, mica schist, quartz-mica schist, and fragments of quartz diorite. The schist inclusions show no contact metamorphism, and their contacts are sharp. Structures within the schist inclusions must have developed before they were included (Fig. 20). A mode is given in Table 1, locality 34. The quartz diorite inclusions are in granodiorite. They are not altered, and some are angular (Fig. 14b).

Origin of Inclusions: The inclusions of schist, quartz diorite, and hornblende gneiss were picked up by the magma during its intrusion. The origin of the dark fine-grained inclusions in the quartz diorite is less certain. The sharp boundaries, the lack of reaction zones or contact metamorphism, and the uniform composition individually and collectively suggest they are inclusions picked up during intrusion. These inclusions may be (1) schist inclusions impregnated with igneous material, (2) segregations within the magma, or (3) rounded fragments of an early phase of the quartz diorite gneiss. Their uniformity of composition, its similarity to the quartz diorite, and their abundance in some rocks and scarcity in others favor fragmentation of an early phase of quartz diorite gneiss.

Injection Gneisses: A few rocks, developed locally and confined essentially to the margins of the complex, are schistose, but their composition is that of quartz diorite or granodiorite. They grade from medium-grained mica schists with thin, closely spaced pegmatite stringers into fine-grained rocks with crumpled mica foliae which bend around quartz and feldspar augen.

The minerals are plagioclase, orthoclase, microcline, quartz, and micas. The accessories are similar to those in the gneisses. Garnet is rare or absent. The foliae which make the bands are less distinct in thin sections than in hand specimens.

The medium-grained rocks seem to be *lit-par-lit* injections of igneous material into schists. The fine-grained rocks resemble sheared gneisses. A mode is given in Table 1, No. 35.

Mylonites: Mylonitic rocks are scattered through the area as thin bands or zones from half an inch to 10 feet or more wide. They are a dull green gray and contain a few recognizable fragments of quartz, feldspar, or mica in a very dense matrix of comminuted materials. Very indistinct and thin banding is recognized, and the bands are thicker and more distinct where the grains are larger. The contacts with adjacent rocks are sharp where observed, and there is no gradation from the mylonite into the wall rock.

Identifiable minerals include clouded plagioclase, angular quartz, strained muscovite and biotite, epidote, and traces of sphene and chlorite. These minerals suggest that quartz diorite and granodiorite were intensely sheared to form the mylonites. In some cases the epidote is euhedral, even where in a dense matrix. Late quartz occurs in veins cutting the bands.

The groundmass is dense, almost isotropic. Banding is the result of slight variations in grain size. Although not siliceous, the rocks are similar in appearance to cryptocrystalline cherts.

MISCELLANEOUS ROCK TYPES: Two rock types of very local and minor occurrence suggest alteration of hornblende gneisses by hydrothermal solutions.

They occur in sec. 24, T. 20 N., R. 18 E., and sec. 4, T. 20 N., R. 19 E., Elmore County, near outcrops of hornblende-biotite gneisses. The first type is a dense, dark-green, hornblende-epidote rock. Epidote is the principal mineral, and hornblende is minor. The epidote occurs in small crystals or as irregular, large poikiloblastic crystals that enclose small epidote and hornblende grains. Joints are filled with quartz and a few pyrite grains. Hornblende is lacking for short distances on either side of the joints.

The second type is a faintly banded, green, epidote rock which contains some pink garnet. The epidote is similar to that in the first type and also encloses small epidote grains and a little pyroxene (?). The garnet grains are ragged and enclose quartz and epidote poikiloblastically.

SEQUENCE OF INTRUSION: As a general rule the normal rock types have been intruded in the following order. The hornblende gneisses are older than the Pinckneyville gneisses, and it is assumed that the hornblende gneisses are derived from igneous rocks. The sequence is: hornblende gneisses, quartz diorite, granodiorite, and then granite. The last three constitute the Pinckneyville complex.

This sequence is the result of a number of continuous and overlapping intrusions and not of sharply separated periods of intrusion. The pegmatites, aplites, and quartz veins also probably overlap one another for they are intimately associated with each rock type of the Pinckneyville complex.

This sequence is based on observations of mutual relations of inclusions, minerals, and structures. The inclusions furnish the most definite evidence. Inclusions of schist in gneiss are common and widespread and indicate that the gneisses are intrusive. Structures within the inclusions show that the schists had been metamorphosed prior to intrusion of the enclosing rock.

Inclusions of hornblende-biotite gneiss in quartz diorite are recognized in a road cut on Alabama highway 9, sec. 4, T. 21. N., R. 19 E., Coosa County.

Inclusions of quartz diorite in granodiorite are found along Elkahatchee Creek east of the bridge on Alabama highway 63, in a ditch near the Alexander City Highway Camp, and in the SW$\frac{1}{4}$ sec. 24, T. 20 N., R. 18 E., Elmore County. These inclusions of quartz diorite are large, blocky, poorly oriented specimens.

The general sequence of crystallization, as shown in the mineral paragenesis, conforms with the general concept of differentiation. The potash feldspar content increases in the younger rocks. There is a suggestion that the biotite composition, as shown by indices of refraction, changes with increase in potash feldspar. The

amount of biotite decreases, and the muscovite content increases in the younger rocks.

The variation in intensity of foliation can be used as a criterion but only in conjunction with other evidence. The intensity of the foliation varies with age but also because of local influence of wall rocks. The younger rocks, as a rule, show a less pronounced foliation than the quartz diorite.

DISTRIBUTION OF ROCK TYPES: The Pinckneyville complex is one large composite intrusive mass. On the geologic map (Pl. 1) no attempt has been made to distinguish the numerous types as a much larger scale would be necessary.

Plate 3 is the cross section and plan of an outcrop in a road cut in sec. 2, T. 21 N., R. 20 E., Coosa County. The frequent repetition of types, the complexity of structures, and the close interbanding of units are represented by a scale which permits the distinguishing without distortion of units as narrow as 4 inches. Such complexity of rock types and structures is not present everywhere on the same scale, and in many outcrops the quartz diorite or the granodiorite is more or less homogeneous. Plate 3 shows, however, a complexity of rock types and structures which is typical of the whole area. The complex is a composite mass in which the individual units are both large and small and result in an interbanding which has much the same pattern regardless of the scale on which it is presented.

Despite this apparent close interbanding of units, certain generalizations on the distribution of petrographic types are possible. Quartz diorite is the typical Pinckneyville rock type. It is most prevalent in the central portion of the area. Field observations suggest that biotite decreases slightly from east to west and from south to north. Granodiorite and granite are more abundant along the margins but occur in the central portion. The intrusive masses are probably not continuous along the strike of the entire surface extent of the complex.

A large area of leucogranodiorite gneiss extends from about Hackneyville to south of Alexander City. (*Cf.* locality 22, Table 1.) The area in which it crops out is covered with a light-colored soil. This rock is exposed in the quarry along Alabama highway 63 just south of Alexander City. Another area of this rock type extends several miles northeast and southwest from Socapatoy. An extensive area of granitic rock crops out along Alabama highway 63 in secs. 17, 20, and 30, T. 22 N., R. 20 E., Coosa County. Pegmatite zones are traced by float for several miles along the strike and underlie many small low ridges.

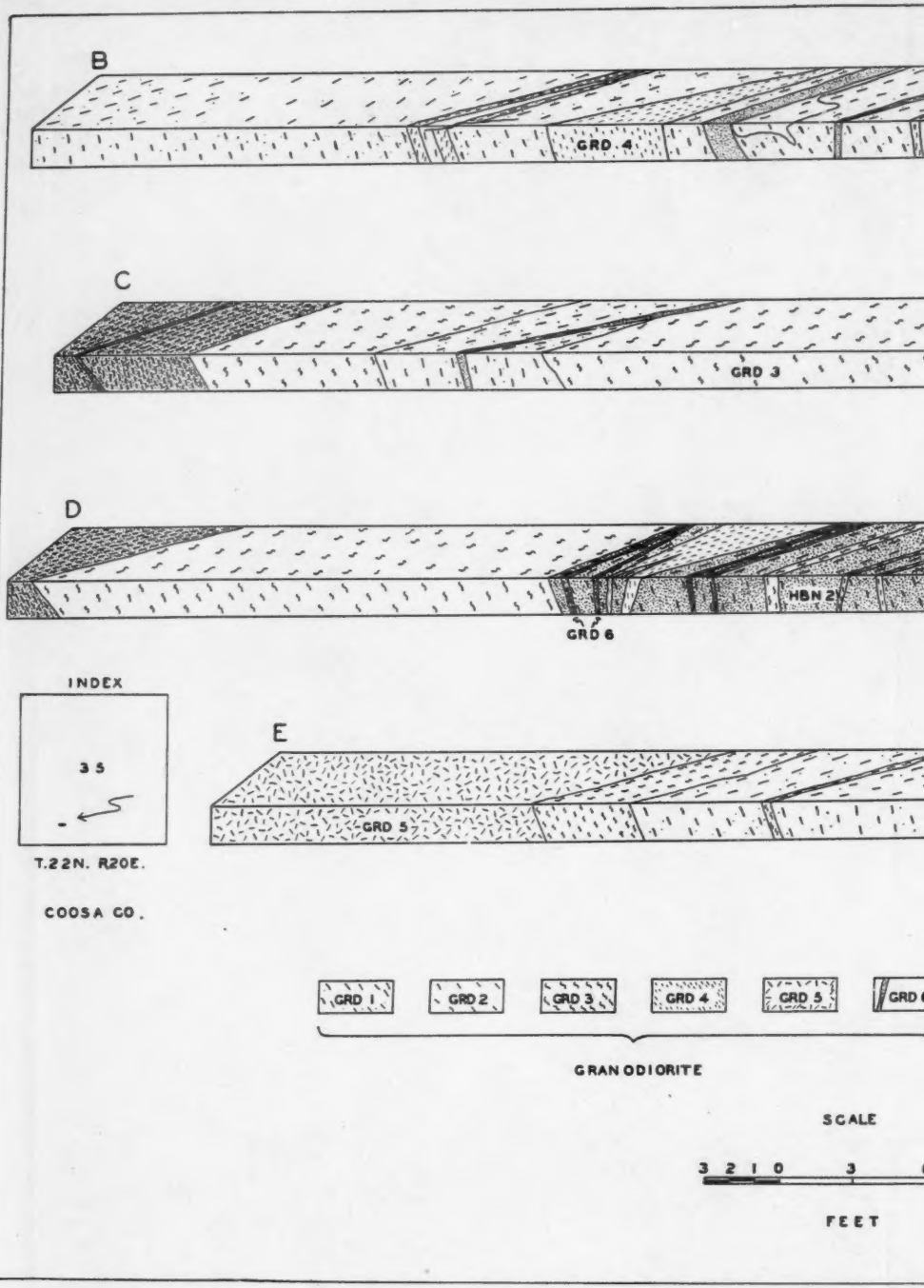
The hornblende gneisses are widespread but always occur in relatively narrow bodies. Exposures are poor because of deep weathering, but some zones are traced for several miles along the strike by float. Hornblende gneisses are more abundant on the margins of the complex. The schist septa decrease in number and size to the southwest. The septa range from the dimensions of those in Plate 1 to septa several feet thick and of unknown length.

STRUCTURES OF THE REGION

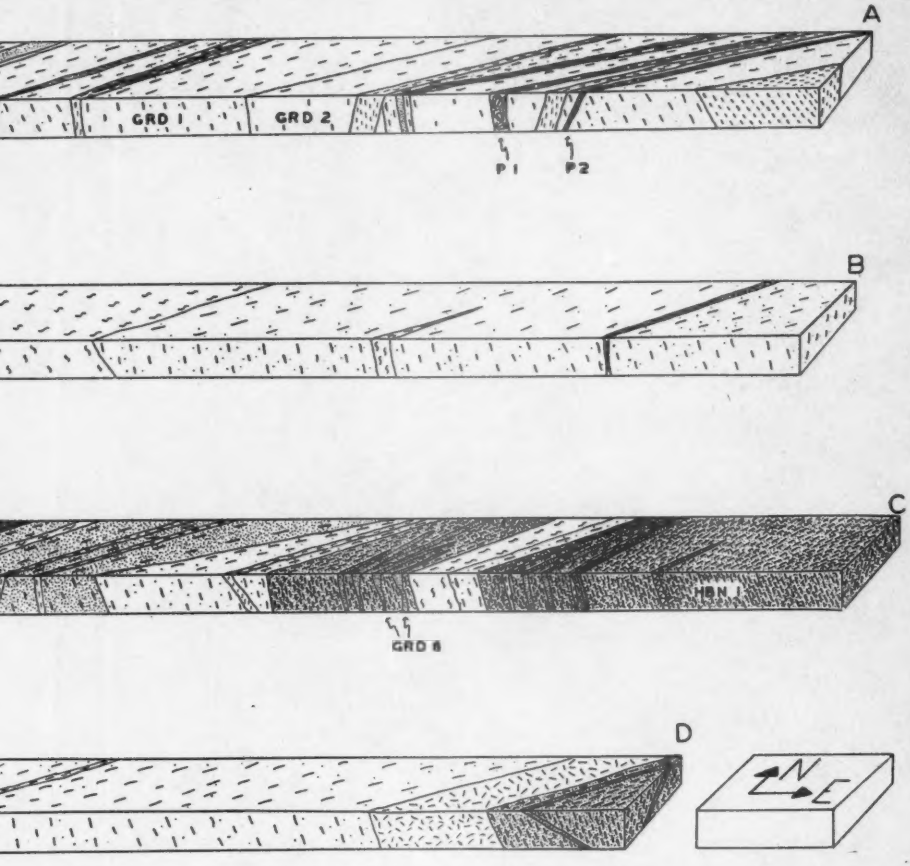
DEFINITIONS OF STRUCTURAL TERMS

FOLIATION: Foliation is a megascopically visible planar structure and is the result of banding or mineral orientation or both.

FLOW CLEAVAGE: This term is used for schistosity and cleavage and is the ability of a rock to



CABINET BLOCK DIAGRAM SHOWING STRUCTURE AN
Sec. 35, T. 22 N., R. 20 E., Co



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URE AND ROCK TYPES IN A SMALL ROAD CUT

. 20 E., Coosa County

part along subparallel planes because of the arrangement of the cleavage directions of mineral grains or of inequidimensional particles.

PRIMARY FLOW LAYERS, FLOW LINES, FLOW STRUCTURES: These are visible because of and are produced by the movement of the magma during its flowage phase.

FRACTURE CLEAVAGE: This term is used to describe incipient close-spaced jointing due to mechanical deformation and not to flowage or reorientation of particles by recrystallization. It can be responsible for small crinkles which represent folded flow cleavage.

LINEATION, LINEAR STRUCTURES: They are the result of orientations of pencil-like or elongate mineral grains or aggregates and the intersections of bedding and cleavage, cleavage and fracture cleavage, and other plane intersections.

CROSS JOINTS: These are joints striking normal to the regional strike and intersecting the foliation and lineation at right angles.

OBLIQUE JOINTS: They intersect the regional structures at any angle.

LONGITUDINAL JOINTS: These joints strike parallel to the regional strike.

b OR *B*: Fold axes, intersections of bedding and cleavage, intersections of flow layers and cleavage, or axes of mica girdles are designated by the co-ordinate direction *b* or *B*. *c* is normal to *b* and to the foliation. *a* is normal to *b* and lies within the foliation.

Strike readings are interpreted as follows: 10 is N. 10 E. or S. 10 W., 95 is S. 85 E. or N. 85 W., 170 is S. 10 E. or N. 10 W., and 180 is the same as zero and is due north or due south.

METAMORPHIC ROCKS

Planar structures.—Except for rude banding the original bedding structures have been destroyed by recrystallization. The banding is due to changes in mineral content or to variations in grain sizes across the cleavage. Fine-grained quartz schists are made up of alternating layers of quartz and mica-rich material. In graphitic schists the bands contain varying proportions of graphite, and in garnet schists the bands are zones of greater concentrations of garnet crystals. The bands are not continuous because of the considerable rock flowage and because lithologic changes across the strike are rapid. The rapid change from one rock type to another makes it difficult to recognize stratigraphic units, and no attempt is made to define such units. Bedding was not observed intersecting the cleavage at a measurable angle.

Since flow cleavage usually parallels bedding it is assumed that the axial planes of the folds are parallel to the cleavage. No large folds were observed because of limited exposures and the lack of detailed work far out into the schists. The axial planes of small folds in bands of quartz and small drag folds in hornblende gneisses dip southeast parallel to the rock cleavage.

The foliation of the metamorphic rocks is a flow cleavage resulting from the parallel growth of platy minerals and from the reorientation of minerals and rock particles into the foliation plane. Flow cleavage is the dominant structure of most of the metamorphic rocks. Rocks lacking platy minerals do not show a good flow cleavage, but planar structures are suggested by faint banding which is probably the original bedding.

Flow cleavage is frequently folded or sheared resulting in a fracture cleavage. Folding of the flow cleavage is visible only under the microscope in some rocks. In others the amplitude of the folds is about half an inch. The crests of these folds are rounded in the coarse-grained rocks and angular in the fine-grained rocks. The coarse-grained rocks usually show larger folds. Fine-grained rocks with fracture cleavage locally weather into "pencils."

Since the fracture cleavage is superposed on the flow cleavage, it is later and prob-

ably represents a later phase of the same deformation which produced the flow cleavage (Cloos, 1937).

Many small faults can be observed in large road cuts or cliffs and in horizontal exposures. Measurable displacements range from a few inches to several feet and account for some of the variations in the strike of the foliation within short distances. Small shear zones occur throughout the area several feet to 15 feet long and a few inches to a foot wide. Within the shear zone, the schists are brecciated or intensely crumpled. The faults and shear zones cut the flow cleavage at acute angles and are later. They may have developed at the same time as the fracture cleavage.

No large faults such as those described by Prouty (1923a; 1923b) in Clay County and by Jonas (1932) in northeastern Alabama were recognized. Recognition of stratigraphic units and detailed structural studies away from the limits of the complex are requisite for their determination.

Joints are abundant in the metamorphic rocks and are grouped in several sets. These sets of joints are not restricted to particular rock types. The joint surfaces are smooth except where deeply weathered. Mineral fillings were not seen. The joints are slightly warped and exhibit greater variations in strike over short distances than do joints in the gneiss.

There are three types of joints: (1) cross or regional tension joints, (2) oblique joints, and (3) longitudinal joints. Not all types occur at one locality, and one may be more prominent at one locality than at another. This is due in part to the direction which the outcrop follows—*i.e.*, joints parallel to an outcrop are generally not so prominent as those at an angle to the outcrop even where those parallel are more abundant.

The cross or regional tension joints are most prominent in the metamorphic rocks and usually occur in all exposures. The joints strike normal to the foliation, and the dip is 80° to vertical.

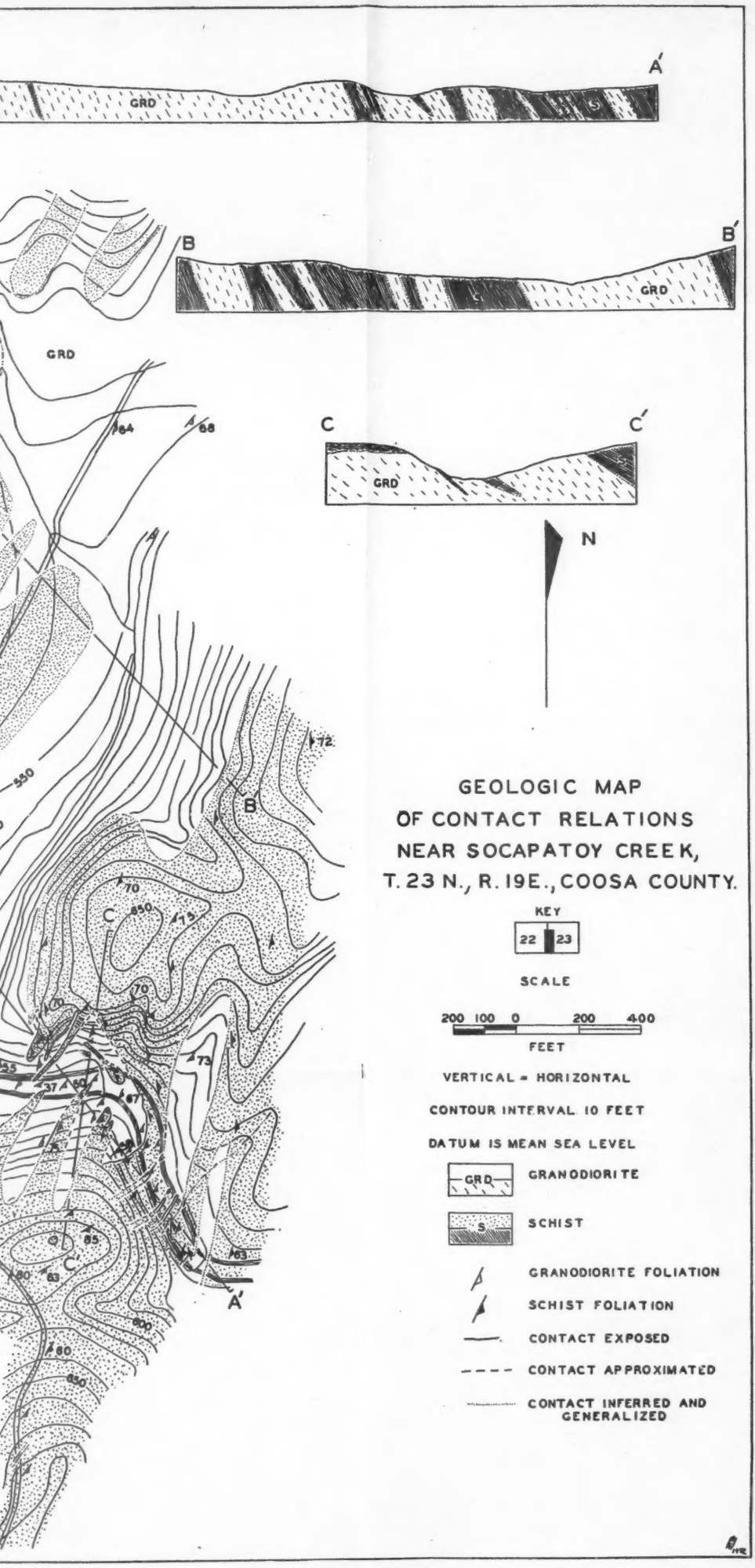
The oblique joints are subdivided into several sets. Two sets are most common and apparently are related to regional structures. One generally strikes northwest and dips southwest. The other set strikes northeast but more easterly than the foliation and dips northwest. Other oblique joint sets are recognized here and there.

The longitudinal joints are not so clearly defined as the others. They seem to fall into three sets: vertical, parallel to the foliation, and dipping gently northwest. The vertical set is most common. It is not certain whether joints parallel to foliation are due to deformation or to weathering. The northwest-dipping set was noted at only one locality near the eastern margin of the complex and may be related to marginal joints in the gneiss.

The above classification is based on field observations. At four localities, 100 or more joint directions were recorded for statistical analysis. The readings represent selected joints, because certain joints in the exposures were small and irregular without any relation to rock structures and probably developed from weathering. The poles of the joint surfaces were plotted on an equal-area net using the lower hemisphere. Concentrations of points within a unit area were counted and contoured similar to the preparation of petrofabric diagrams (Knopf and Ingerson, 1938, part II; Haff, 1938). D 61, D 62, D 63, and D 67 represent, statistically, joints in the schists (Pl. 8).

GEOLOGIC MAP OF CONTA

Socapatoy Creek



OF CONTACT RELATIONS BETWEEN SCHIST AND GRANODIORITE

ocapatoy Creek, secs. 22 and 23, T. 23 N., R. 19 E., Coosa County

D 61 shows the joint pattern in a cherty, brecciated quartzite exposed in a quarry in sec. 10, T. 21 N., R. 20 E., Coosa County. A nearly vertical cross joint set and a longitudinal set parallel to foliation and bedding are indicated in the diagram.

D 62 was constructed from readings gathered in a series of road cuts in sec. 35, T. 21 N., R. 19 E., Coosa County, a quarter of a mile east of the main contact. Two oblique joint sets and a less prominent cross joint set are shown in the diagram.

D 64 presents the joints in the schist immediately adjacent to the granodiorite intrusive along Socapatoy Creek (Pl. 4). Cross joints and vertical longitudinal joints appear to make the two prominent sets, but the schists are considerably disturbed here by the granodiorite (D 65), and the relations of the joints to rock structures is not clearly defined.

D 67 depicts the joints in the schist along Elkahatchee Creek for 300 feet downstream from the contact (Pl. 5). The joints are clearly defined in the field, as shown by the excellent concentration of the poles in the diagram. The principal sets are the cross joint set, a southwest-dipping oblique set, and a northwest-dipping longitudinal set which corresponds to the marginal joints in the gneiss.

Linear structures.—The fold axes lie in the flow-cleavage plane and pitch 10°–40° NW. or SE., averaging 10°–20°. The pitch of fold axes approximately parallels the pitch of linear structures in the gneiss complex.

Intersections of flow cleavage and bedding were not observed. The intersection of flow cleavage and fracture cleavage lies in the plane of flow cleavage approximately parallel to fold axes.

Linear structures on flow-cleavage surfaces are expressed by elongated platy minerals, spindle-shaped pods of quartz and garnet, and striations. Both biotite and muscovite grow in long blades parallel to *b*. Coarse mica schists are composed of large, wavy, relatively thick mica crystals which have a long direction subparallel to *b*. In finer-grained rocks the mica flakes are small and thin.

Many quartz pods are drawn out into spindle-shaped bodies surrounded by mica. The long axes of the spindles parallel *b* on flow-cleavage surfaces, and the surfaces are knobby. Garnet spindles develop with similar orientations.

Striations parallel to *b* seem to be due to slippage along the flow cleavage surfaces.

Folds.—All folds observed are small. In the schists such folds approach isoclinal types but are more open in more competent beds. Large regional folds are indicated by the northwest dip of the foliation in Elmore County several miles east of the Pinckneyville complex.

Both shear and bent folds occur. Shear folds are to be expected in a region with a well-developed flow cleavage (E. Cloos, 1937) and are shown by the small folded quartz lenses with thick crests and thin limbs. Bent folds include the crinkled flow cleavage and fracture cleavage. These folds are later than the development of flow cleavage.

Distribution of structures.—Flow cleavage is prominent throughout the area. Its strike ranges through 360°, but most readings lie between N. 5° W. and N. 45° E. The dip generally ranges from 50° SE. to vertical, but dips as low as 25° SE. or reversals to steeply northwest are recognized.

Linear structures are widely distributed but are not present in all exposures. Fracture cleavage is more common in the belt of cassiterite-bearing pegmatites in schist

between Goodwater and Rockford on the west side of the complex than on the east side. Crinkling of the flow cleavage and the development of fracture cleavage in this belt is so intense at many places that the earlier flow cleavage is obliterated.

IGNEOUS ROCKS

General statement.—The structures of the igneous rocks developed in three phases in the flowage phase or during emplacement of the magma, in the transition phase or after emplacement but prior to complete crystallization and cooling, and in the solid phase or after consolidation of the magma. The structures of these three phases are not everywhere distinct, and at many places they occur together. Structures of a later phase are superposed on earlier structures. The structures of the flowage phase are primarily the result of internal forces or magmatic flow; the structures of the transition phase are the result of both internal forces and external forces or regional deformation; and the structures of the solid phase are primarily the result of external forces acting on a solid body.

Structures of the flowage phase.—Planar-flow structures due to the arrangement of mineral grains, aggregates of grains, and inclusions are well developed at many places throughout the area. Despite their widespread occurrence, structures of the flowage phase are less prominent than those of the two later phases and are masked by the later structures.

Biotite and muscovite show the best flow structures, particularly biotite because of its dark color. The micas occur as individual grains or more commonly as aggregates of subparallel flakes. The grains and aggregates are usually slightly elongated. The intensity of the foliation varies with their degree of parallelism.

Coarse biotite flakes also accumulate in bands 6 inches to several feet long and a fraction of an inch wide. These are uncommon but widely distributed and always parallel foliation.

Feldspar crystals are not readily oriented during flow because they are nearly equidimensional. In rocks with a distinct foliation, the largest face of each feldspar crystal tends to parallel foliation. As the foliation becomes more pronounced the feldspars show protoclastic granulation and are in better-developed augen. The feldspar augen are surrounded by mica and quartz.

At some places the flow structure in the younger rocks follows the contact adjacent to it, but a short distance away it is not influenced by the wall rock (Fig. 13a). Where the foliation is more pronounced, flow structures are observed only near irregularities of the contact (Fig. 13b) or around inclusions (Fig. 14c).

The dark fine-grained inclusions in the quartz diorite are almost always platy and are about parallel to the foliation (Figs. 14a, c). Inclusions of schist usually parallel foliation. On the west wide of the complex along Socapatoy Creek (Pl. 4) about half the inclusions in the granodiorite parallel foliation, and the others are arranged at random. The intrusion is close to its roof here, and the rate of flow probably differed from place to place and followed several directions.

The orientation of quartz diorite inclusions in granodiorite is poor at some places (Fig. 14b) because of the irregular shape of the fragments, but groups and chains of inclusions tend to align themselves parallel to the regional trend. The shape and

arrangement of the inclusions suggest brecciated zones. Other examples of quartz diorite inclusions in granodiorite show a much better alignment (Fig. 14c).

Structures of the transition phase.—The foliation and lineation which developed in the transition phase are visible chiefly because of mica orientations. They are similar to those of the flowage phase but are more uniform and are related to regional

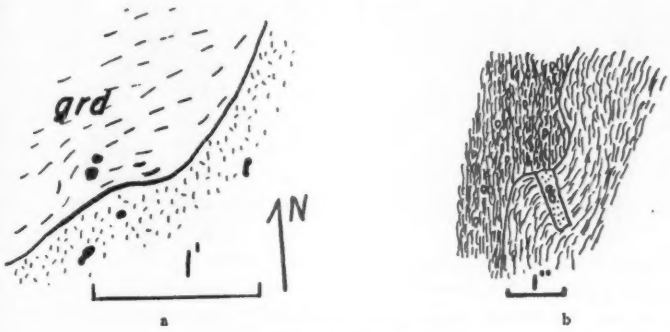


FIGURE 13.—*Flow structures*

a. Flow structure in granodiorite shown by orientation of biotite. Influence of contact is local; Socapatoy Creek, Coosa County.

b. Mica orientations and quartz-filled feather joint in quartz diorite; Elkahatchee Creek, Tallapoosa County.

structures rather than to local contacts. The long dimensions of the micas are about parallel to a horizontal or nearly horizontal axis which is *b*. This arrangement of the micas produces a lineation in *b*. Where the short dimensions of the micas are arranged somewhat uniformly around and normal to *b*, the foliation is indistinct, and several faint planar structures at angles to one another are recognized. Where the foliation is distinct many of the mica plates and their short dimensions are about parallel and determine the foliation. The intersection of the mica plates with northeast-trending surfaces also produces a lineation.

A few of the dark, fine-grained inclusions in the quartz diorite are spindle-shaped rather than platy, and they are linear elements. These spindle-shaped inclusions are oriented parallel to the lineation of the micas.

Near many external and internal contacts the foliation is strongly gneissic or schistose, and border gneisses are developed. Augen structures in the border gneisses are pronounced, and the feldspar grains are drawn out into long lenses as a result of protoclastic deformation and are surrounded by micas and quartz. The quartz occurs as thin wavy patches about parallel to the foliation.

Some of the gneissic structures may have developed during intrusion as a result of greater viscosity and frictional drag of the crystals in the magma along the walls of the intrusive. However, most of the border gneisses probably developed in the transition phase as zones of adjustment to regional deformation. These zones may have started to develop in the flowage phase. Examples of border gneisses considered to be the result of increasing viscosity and greater friction along the walls are described by Waters and Krauskopf (1941), Balk (1937), Barrell (1921), Adams and

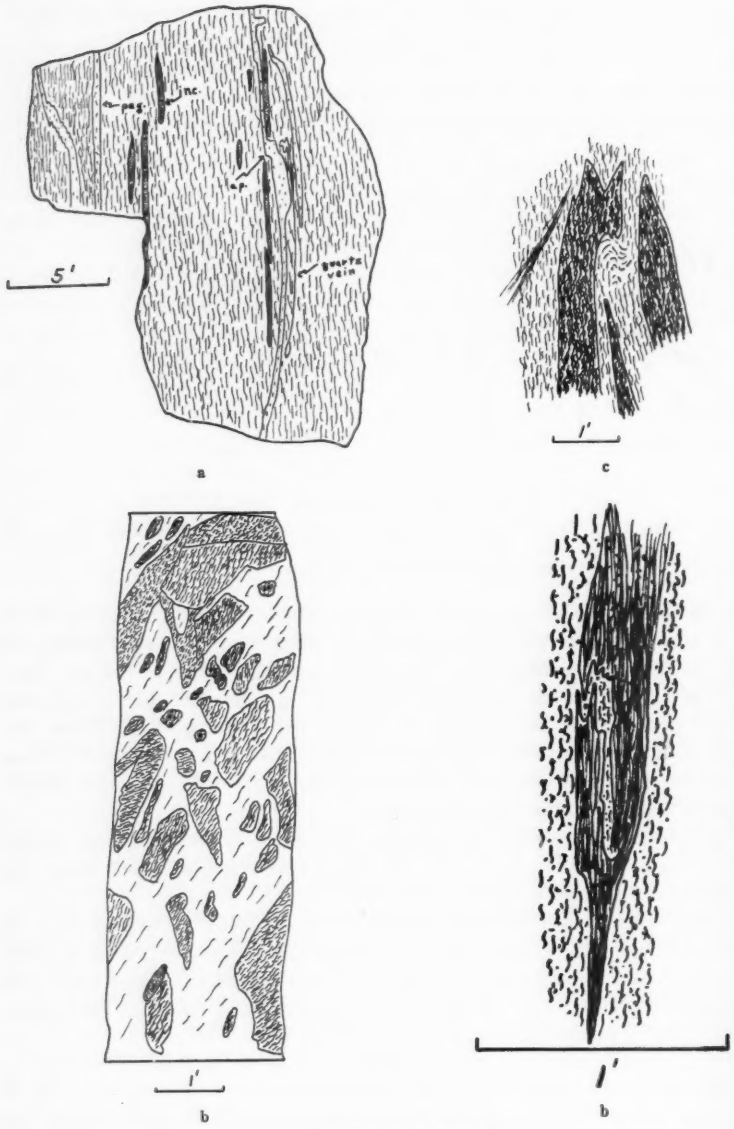


FIGURE 14.—*Inclusions*

a. Inclusions in quartz diorite; Elmore County.
b. Quartz diorite inclusions in granodiorite. Structures in granodiorite generalized; sec. 23, T. 23 N., R. 21 E., Tallapoosa County.
c. Quartz diorite inclusions in granodiorite; Elkahatchee Creek, Tallapoosa County.
d. Folded schist inclusion in quartz diorite; sec. 4, T. 20 N.,R. 18 E., Elmore County.

and Barlow (1910), and many others. Berkey and Rice (1919) have attributed structures similar to these border gneisses largely to regional deformation prior to complete consolidation rather than internal convection in the magna.

Marginal fractures related to contact zones have been described as marginal thrusts (Balk, 1937). Marginal fractures, showing no observed displacements, are developed in the gneiss along Elkahatchee Creek (Pl. 5) for a maximum distance of 800 feet in from the contact. They dip northwest away from the contact into the Pinckneyville complex. Fractures of this type are not developed within the main body of the complex.

Small flexures of the flow structures locally pass into small fractures which at some places are filled with quartz (Fig. 15a). Within a few inches of many of these fractures the foliation is dragged as much as 45° away from its normal trend. Here and there inclusions are fractured, and parts of the inclusions are displaced along the fractures. The fractures die out almost immediately on either side of the inclusions (Fig. 15b).

Wegmann (1932) describes a structure due to stretching of competent material surrounded by incompetent material in which the competent layer yielded by rupture and the matrix by flowage. This structure is termed boudinage, and the parts which developed by rupture are boudins. Some schist inclusions show boudinage structures (Fig. 15c). These inclusions apparently were pulled apart after their incorporation in the magma but prior to complete solidification. The flow structures of the enclosing rock tend to follow the outline of the boudins. This structure indicates stretching in *b*.

Structures of the solid phase.—One of the prominent results of forces acting on a solid body is fracturing or jointing. Joints are well developed in the Pinckneyville, and considerable attention was given them. Displacements along some fractures where contrasting bands and dikes were offset resulted in recognizable displacements. Some rocks yielded by intense shearing and mylonites developed. Many dikes and veins followed the fractures and are included under structures of the solid phase.

JOINTS: Feather joints or tension joints (E. Cloos, 1922) produced by movements with negligible displacements are distributed throughout the Pinckneyville complex. Many of these feather joints are arranged en echelon in zones along regional joints (Fig. 16). Others occur singly where slight differential movements between two rock types produced short open fractures. Chlorite and epidote fill many of the feather joints in the zones along regional joints, and quartz generally fills the other feather joints.

Regional joints are developed everywhere in the complex and are grouped in sets similar to those in the schists. Joint surfaces are irregular and smooth, and some are slickensided. The joints of any one set show less range in strike than those in the schists.

A small portion of the oblique and cross joints show bleached zones 2–5 cm. wide, which weather to small rounded ridges. Within the zones of bleached material the feldspars are cloudy and slightly silicified. Biotite is chloritized, and the chloritization extends beyond the limit of the clouded feldspars. Some joint surfaces are

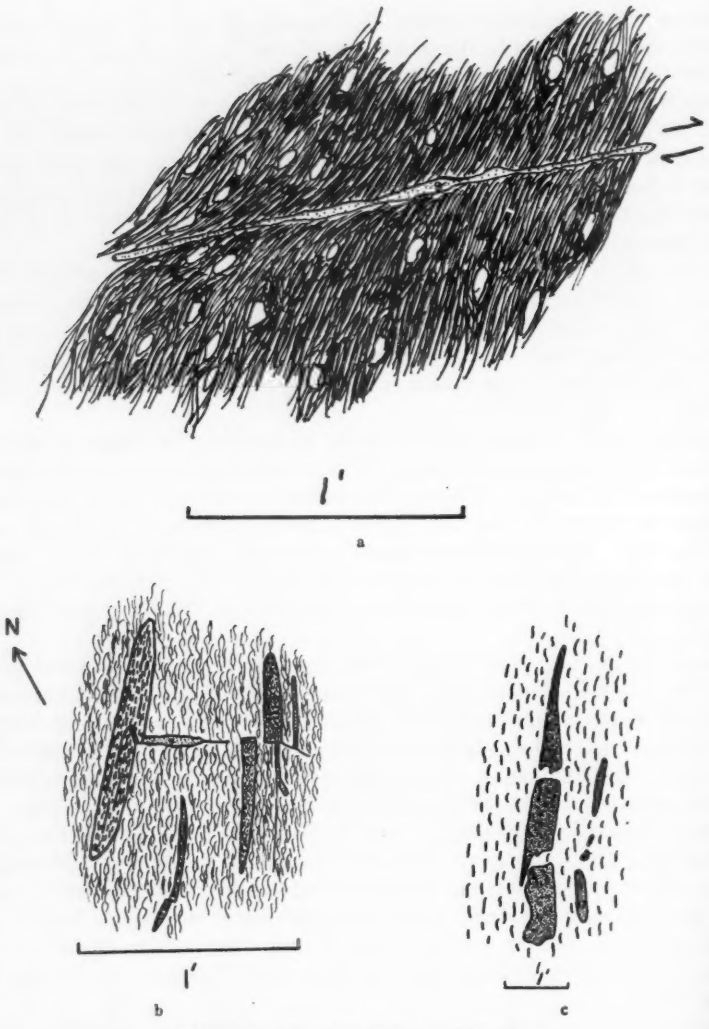


FIGURE 15.—*Structures of the transition phase*

a. Flexure and quartz-filled fracture in granodiorite; sec. 36, T. 22 N., R. 20 E., Coosa County.
b. Faulted inclusions and quartz-filled fractures; Elkahatchee Creek, Tallapoosa County.
c. Boudinage structure of inclusion in granodiorite. Flow structures in granodiorite follow fragments; Elkahatchee Creek, Tallapoosa County.

coated with chlorite and small amounts of epidote. The width of the filling varies (Fig. 16).

Joints are grouped under cross joints, oblique joints, and longitudinal joints.

Cross joints are the most pronounced and are distributed throughout the complex. They are vertical or almost vertical and are about normal to *b*.

Oblique joints are more numerous than other joints, but not all of them are related to other structures. Two sets of oblique joints similar to two sets of oblique joints in the schists are recognized. One set strikes between 145 and 170 and dips southwest at moderate to steep angles. The angle between the strike of the joints and

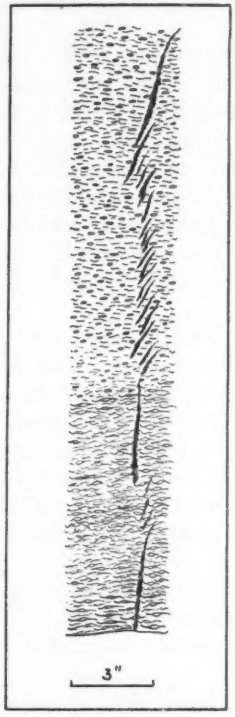


FIGURE 16.—*Feather joints*

Along a regional joint filled with chlorite and epidote. Note discordant contact of gneiss, sec. 2, T. 21 N., R. 20 E., Coosa County.

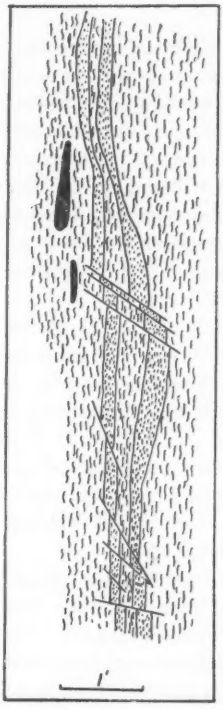


FIGURE 17.—*Small faults in granodiorite*

Sec. 4, T. 21 N., R. 20 E., Coosa County.

the strike of the foliation is 35°–40°. The second set strikes 85 to 115 and dips moderately to steeply north. The angle between the strike of the joints and the strike of the foliation is 35°–45°. These two sets fall in the tension quadrant (H. Cloos, 1921).

Two minor oblique sets appear in a few places which are in the compression quadrant (H. Cloos, 1921) and may be the conjugate joints of those in the tension quadrant. One of these two minor sets is similar to the glide joints described by Balk (1927). The joints are closely spaced and small, strike 10°–20° east of the foliation, and dip moderately southeast. The surfaces are slickensided and coated with mica.

The second minor set is more widely distributed but is not prominent. It strikes about 10°–20° north of the foliation and is steep to vertical. Many granite dikes

follow this direction in the Pinckneyville complex, and Pratt (1908), Weed (1911), and Laney (1917) have mentioned the more northerly trend of sulfide veins in the southern Appalachians making angles of 10°–15° with the regional planar structures.

Longitudinal joints are common loci for dikes. The joints are rough and are parallel to or cut the foliation. Marginal joints are included with the longitudinal joints.

Several statistical analyses of joints in the gneiss were made (Pl. 8). D 63 shows the joint pattern in the gneiss just west of locality D 62. Cross joints dominate with some southwest-dipping oblique joints. This pattern differs from that in the schist.

D 65 represents the joints in the granodiorite at the locality of D 64. D 65 and D 64 include only joints exposed along the stream, and the patterns in the two diagrams are dissimilar. Field observations suggest the presence of cross joints and vertical oblique joints. These two sets are also indicated in the diagram D 65.

D 66 represents the joints in gneiss on the east side of the complex in southern Coosa County (Pl. 8). Northwest-dipping oblique joints are prominent with some cross joints.

D 68, D 69, and D 70 represent joints along Elkahatchee Creek (*cf.* Pl. 5, D 67). D 68 includes readings made from the contact for 250 feet west; D 69 for 250 feet upstream from the west edge of the area of D 68; and D 70 the remainder of the mapped outcrop. Cross joints are prominent in all three diagrams although the dip is variable. D 68 and D 69 show gently northwest dipping longitudinal joints. This set also occurs in the schist adjacent to the contact and is the marginal joint set. It disappears about 400 feet upstream from the contact. The southeast-dipping longitudinal set of D 67 is absent in D 68 and D 69, and only cross joints are common to D 67 and D 70.

D 71 is made from readings on a few joints in each of many small exposures of gneiss in Elmore County. Cross joints are most prominent with a minor vertical oblique set also developed. These joints are most commonly observed and are those which influence weathering most.

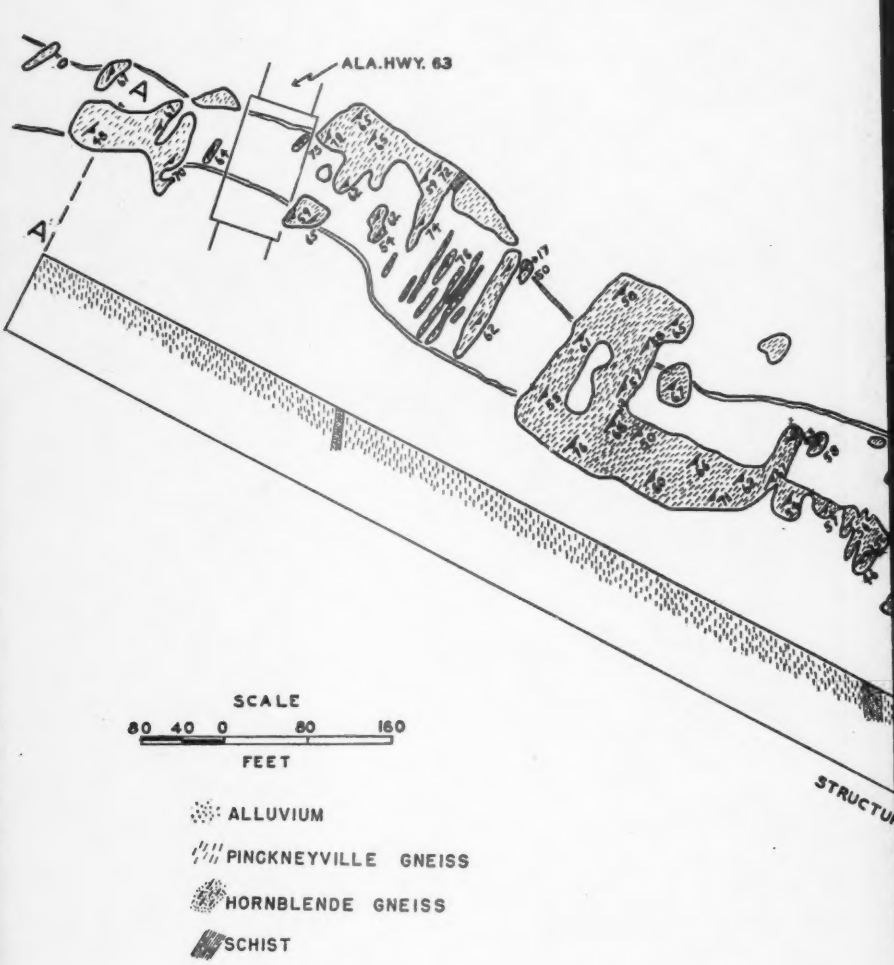
DISPLACEMENTS: Mylonite bands half an inch to 10 feet thick parallel the regional strike. These mylonites represent zones along which the rocks yielded to regional stresses by intense shearing movements after consolidation. These zones may have begun to develop in the late stages of the transition phase. Measurable displacements were not observed. Hydrothermal solutions evidently followed these zones because epidote pseudomorphs after feldspar and altered biotite are common constituents within the zones.

Many small faults whose displacements are 10 inches or less cut the rocks of the complex. These faults are generally recognized only where they cut thin, alternating bands of light and dark gneiss or where dikes are offset (Fig. 17). Because many of the dikes are parallel to the regional strike, most of the faults are in longitudinal zones. The faults which cut dikes die out short distances on either side of the dikes. At some localities the foliation is dragged along the faults.

The faults dip steeply and are parallel to prominent joint directions. Many faults follow oblique joint directions and are equally distributed between northwest and southwest directions. In any one outcrop one direction commonly predominates (Fig. 17). Faulting in these directions indicates a lengthening parallel to *b*.

BUL

PLAN



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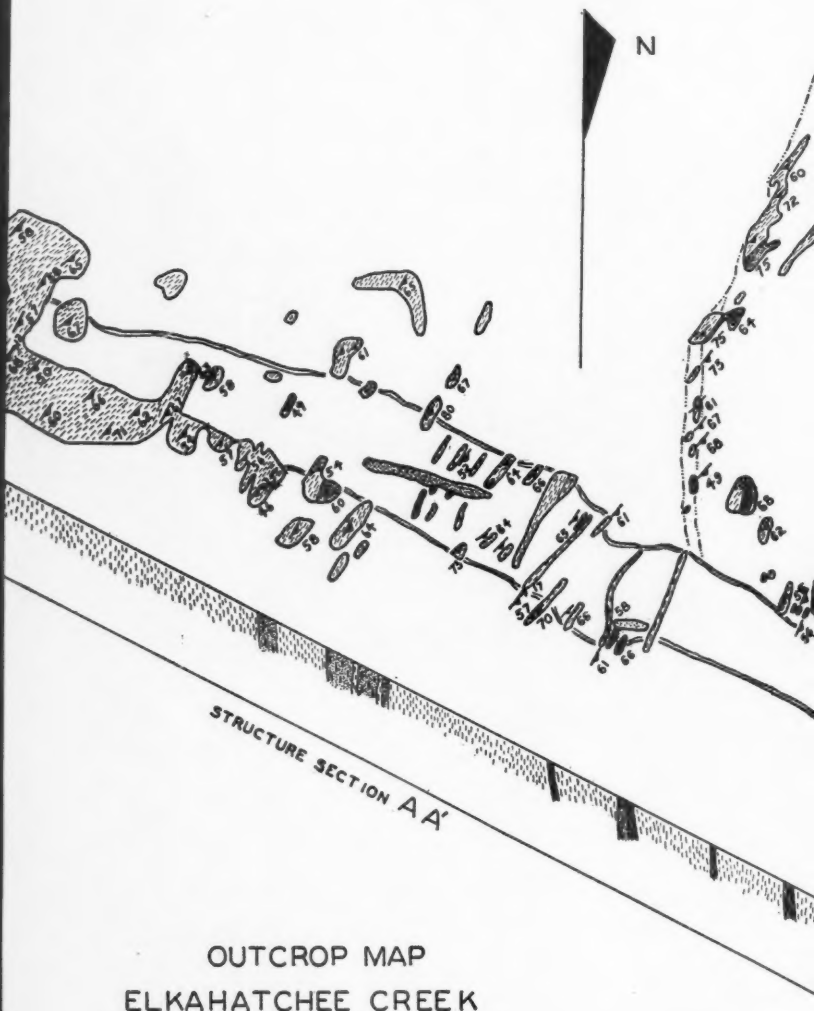
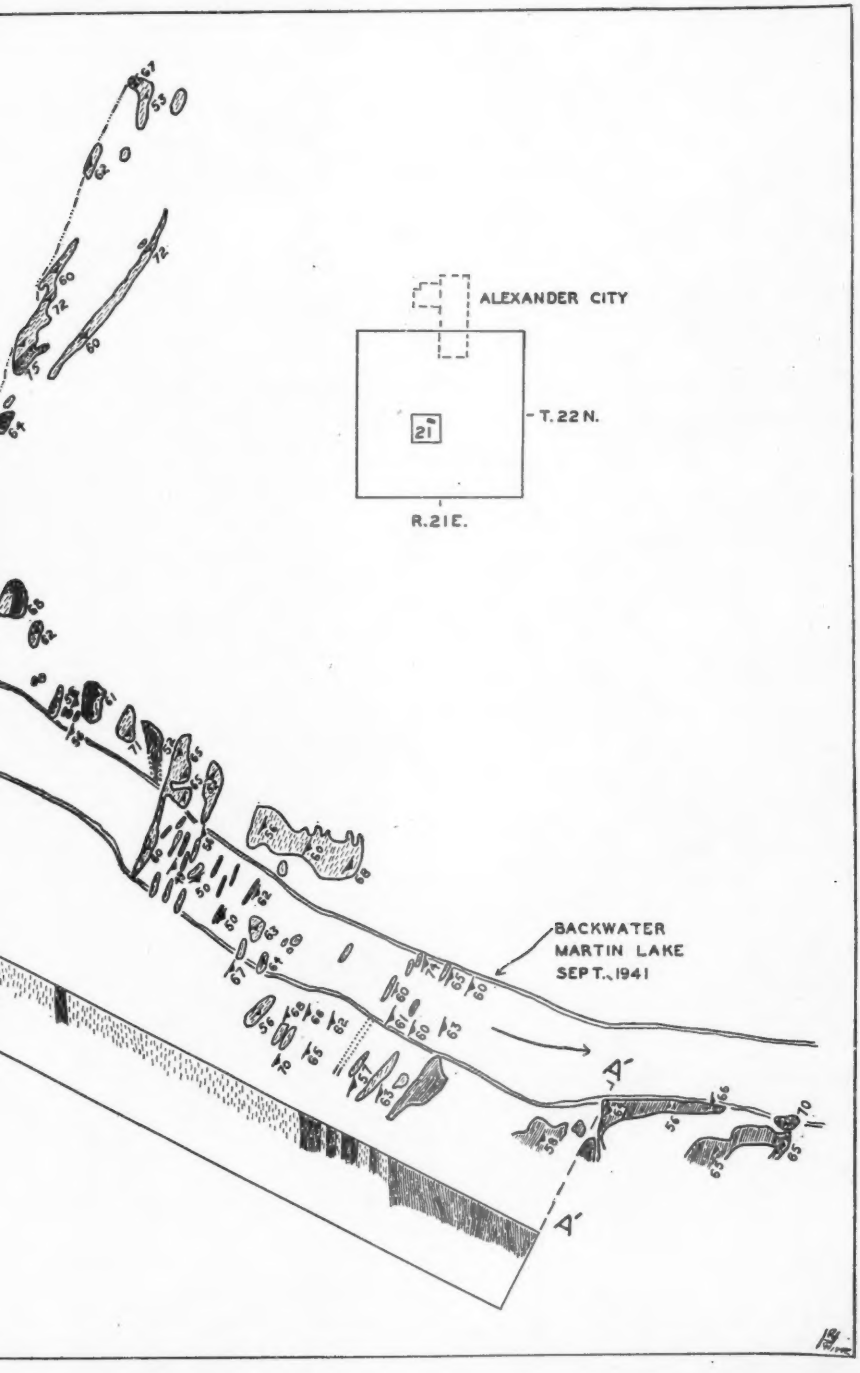
PLANE TABLE SURVEY-1941-H.R.G.&R.P.

OUTCRO
Includes main external contact of Pinckne



OUTCROP MAP OF PART OF ELKAHATCHEE CREEK

les main external contact of Pinckneyville complex. Water level of Martin Lake 20 feet below no



et below normal September 1941

DIKES: Most of the dikes are fine-grained porphyritic and nonporphyritic granites. Some are discordant (*i.e.*, cut the wall-rock structures), generally at an angle of 10°–15° to the foliation. A few make larger angles with the foliation. Cross joints are rarely filled with dike material.

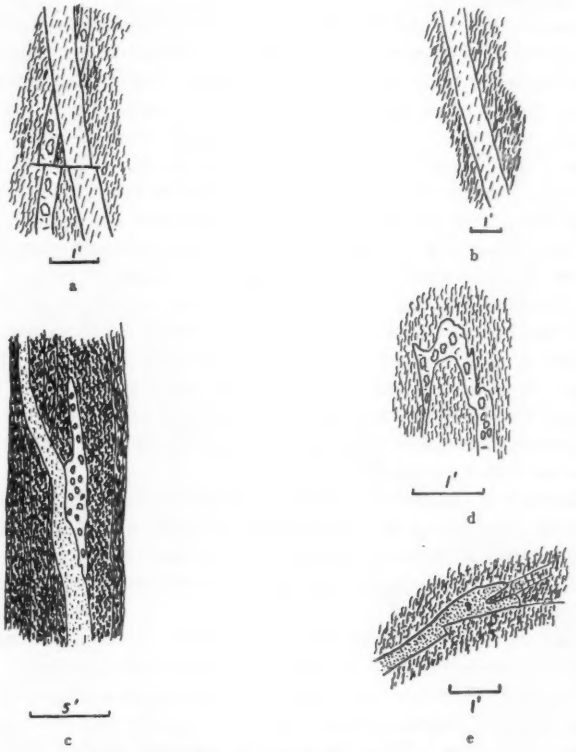


FIGURE 18.—*Dikes, pegmatites, and quartz vein*

a. Fissure pegmatite cut by granite dike and both offset by small fault; sec. 11, T. 20 N., R. 17 E., Elmore County.
b. Discordant conformable dike; sec. 30, T. 20 N., R. 18 E., Elmore County.
c. Discordant conformable dike and concordant pegmatite; sec. 30, T. 20 N., R. 18 E., Elmore County.
d. Ptygmatic fold in pegmatite; sec. 6, T. 19 N., R. 18 E., Elmore County.
e. Quartz vein in faulted zone; sec. 5, T. 20 N., R. 18 E., Elmore County.

The foliation is well developed and in most dikes is conformable (*i.e.*, parallel to the foliation of the wall rock) (Fig. 18). At a few localities the dike foliation parallels the contact.

Most of the dikes are 3–6 inches thick and on flat outcrops extend for distances up to 200 feet without any variation in width or direction. The contacts are sharp but rough. A few dike contacts show small delicate ptygmatic folds. Several dikes are folded by shearing along the foliation but show little granulation or crushing.

PEGMATITES AND APLITES: Pegmatites are common and occur in all rock types. They range in thickness within the complex from a fraction of an inch to 2 feet.

Pegmatites in the schist belt between Goodwater and Rockford attain thicknesses of at least 25 feet. Foliation is not well developed in the pegmatites. Muscovite crystals, however, are bent and enclose feldspar crystals. In a few pegmatites quartz occurs as long narrow rectangular grains. Many of the feldspar crystals are fractured and granulated.

The pegmatites are divided into three types: (1) segregation pegmatites, (2) folded pegmatites, and (3a) regular and (b) irregular fissure-filling pegmatites. These groups are similar to those described by Chapman (1941). Segregation pegmatites are irregular masses of varying sizes (Fig. 18c) but are usually about 1–2 feet long. Some grade into the surrounding rocks, and many contain quartz centers. The crystals are about 5 times as large as those making up the surrounding rock.

Ptygmatically folded pegmatites are most common in the more strongly foliated rocks (Fig. 18b). The plunge of the fold axes is gentle and approximately parallel to other linear structures. These folded pegmatites grade into segregation pegmatites and are thought to have developed during the transition phase.

Regular fissure pegmatites (Fig. 18a) are the most abundant, and many continue for several hundred feet with only slight variations in thickness and strike. They usually show a rude foliation due to muscovite and biotite orientations and to development of feldspar augen. Most of them strike northeast and parallel or dip more steeply than the foliation. A few are oblique to the foliation. The regular fissure pegmatites are earlier than some dikes and faults (Fig. 18a) and are probably later than the folded pegmatites.

The structures of the irregular fissure-filling pegmatites are essentially like those of the regular pegmatites, but they pinch and swell along their strike (Fig. 19). Quartz centers in the swells are common. They are not so strongly foliated as the regular pegmatites. No crosscutting relations were noted between the regular and irregular pegmatites.

Aplites are much less abundant than pegmatites. Their structures and relations to normal rock types are similar to those of the pegmatites (Fig. 14a).

QUARTZ VEINS: There are three types of quartz veins: segregations, folded veins, and fissure fillings. Segregations are the least common and fewer in number than segregation pegmatites. They are irregular rounded patches rarely exceeding 8 inches in diameter, and more commonly are 2 to 3 inches.

Folded quartz veins are also uncommon. They are folded by shearing along foliation in the strongly foliated gneisses. The limbs are less than an inch thick, and the amplitude of the folds is about 2 feet.

There are two types of fissure-filling quartz veins: gash or feather joint fillings and long regional joint fillings. Most of the feather joint fillings are 1–3 inches long and less than an inch thick (Figs. 13b, 15b). A few are 6 inches thick.

Quartz veins in regional joints follow longitudinal and oblique joints. A few quartz veins are in small fault zones (Fig. 18e). Some show flexures and offsets (Fig. 19d). Many veins 10 feet or more long are less than an inch thick and pinch and swell along their strike. Shorter veins average about 1½ inches.

Structures in inclusions.—All inclusions are at least slightly foliated, and their foliation is about parallel to the regional foliation at most localities. The foliation of the inclusions in quartz diorite and of the quartz diorite inclusions in granodiorite is not

well developed. One inclusion (*cf.* Fig. 7b, D 15, D 16) has a faint foliation parallel to the regional foliation. The biotite flakes in this inclusion are folded and sheared, whereas the micas in the enclosing rock are not sheared.

Schist inclusions usually show distinct foliation, and bedding is recognized in some. Sandy layers surrounded by micaceous material show boudinage structures in several

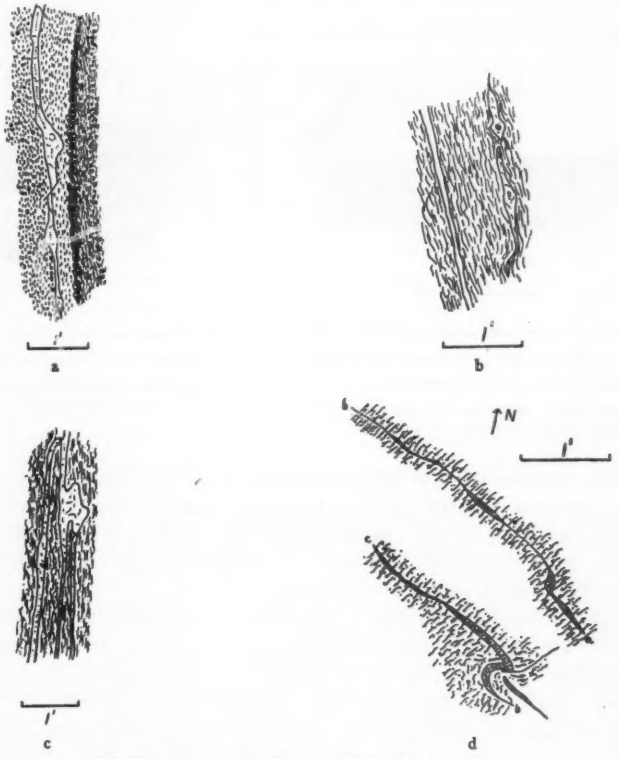


FIGURE 19.—*Pegmatites and quartz vein*

a and b. Irregular fissure pegmatites; Elkahatchee Creek, Tallapoosa County.
c. Pegmatite in quartz diorite gneiss.
d. Quartz vein in cross joint; sec. 5, T. 20 N., R. 9 E., Elmore County.

inclusions (Fig. 20). The boudinage structures do not pass into the enclosing rock and developed prior to their incorporation.

Distribution of structures.—In the northern part of the area, the foliation readings range between 10 and 50 but are generally 25 to 35. The foliation dips 50° SE. to vertical and averages 70°. In the southern part of the area the average strike of the foliation is 35 to 40 along the eastern margin, 30 to 35 in the central part, and 15 to 25 on the western side. The dip is steep to the southeast on the eastern and western margins and steep to moderate to the northwest in the central portion.

In general the foliation becomes more prominent from west to east and from north

to south and is less intense in the younger rocks excluding the development of the border gneisses.

Lineation is widely distributed. The linear structures in the complex are approximately parallel to those in the wall rocks and usually pitch 10°–20° NE. or SW. Many of the linear structures in the northern part of the complex pitch southwesterly; many in the southern part of the complex pitch northeasterly.

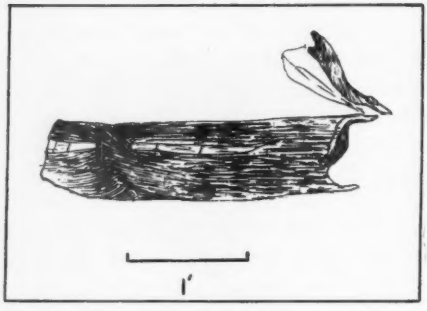


FIGURE 20.—*Boudinage structure in schist inclusion*

In granodiorite. Boudinage structure does not pass into granodiorite; Socapatoy Creek, Coosa County.

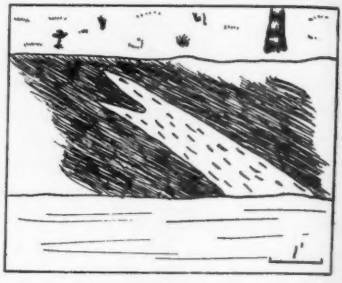


FIGURE 21.—*Granodiorite dike in schist*

Southeast-dipping surface of roadcut; sec. 16, T. 23 N., R. 20 E., Coosa County.

Border gneisses are developed principally along the eastern margin in Coosa and Elmore counties. North from about where the eastern boundary swings to the north in Tallapoosa County and continues into Clay County, there are almost no border gneisses, even immediately adjacent to the main boundary. Border gneisses are rare on the western side of the complex.

Locally, where the foliation is particularly well developed within the complex, the rocks are not so gneissic or schistose as the border gneisses. The local development of intense foliation may be due in part to strong frictional drag along an internal contact. Such local zones of intense development of the foliation are thought to represent zones of material which lubricated adjustments to regional forces within the complex. These adjustments probably continued from the transition phase into the solid phase.

Contacts.—The petrographic boundaries between the igneous components of the complex are both well defined and gradational. Those between quartz diorite and granodiorite are usually distinct, but many contacts between types of quartz diorite or granodiorite are gradational within a narrow zone.

Petrographic boundaries between the rocks of the Pinckneyville complex and the metamorphic wall rocks are always sharp (Fig. 21) even where thin seams of igneous material are injected into schist (Fig. 22).

Most of the small units within the complex are concordant and conformable. Although the contacts of larger units are exposed at only a few places, they also are probably concordant and conformable since the structures of the larger units are the same as those of the small units.

At a few places discordant contacts are disconformable within a narrow zone parallel to the contact and become conformable away from the contact. A small outcrop in Coosa County shows a light-colored, fine-grained quartz diorite in contact with a dark, coarse-grained quartz diorite. The contact strikes 22 and dips 75° NW. The foliation of the dark rock strikes 45 and dips 70° SE. The foliation of the light-colored rock parallels the contact in a zone 2 feet wide. Two feet away from

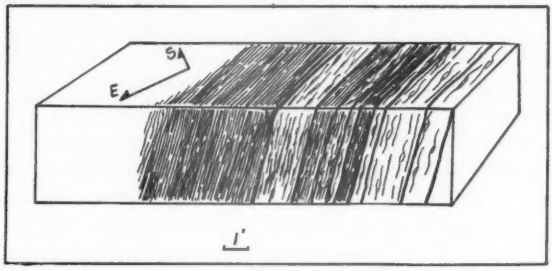


FIGURE 22.—*External contact between schist and gneiss*
Schist to left and gneiss to right; Elkahatchee Creek, Tallapoosa County.

the contact the attitude of the foliation of the light-colored rock is essentially the same as that of the dark.

Dike contacts are both concordant and discordant. Concordant dikes are conformable (Fig. 17). Discordant dikes are commonly conformable (Figs. 18c and 18a), but some are disconformable.

The main contact of the Pinckneyville complex with the metamorphic rocks, and the local contacts of schist septa in the complex or of igneous bodies outside the complex with metamorphic rocks, are essentially concordant and conformable. Intrusion of the Pinckneyville complex was controlled to a great degree by pre-existing structures so that the contacts of the igneous bodies essentially parallel wall-rock structures. Because of the tabular form the contacts are discordant and disconformable generally only at the terminations of the tabular bodies. The terminations of small dikes in schist (Figs. 21, 27) illustrate this type of structure, and the wedge-shaped termination of the complex to the north is a repetition on a regional scale of the termination of the small dikes. Locally at the end of schist septa, the foliation of some granodiorites is conformable and diverges slightly from the regional trend.

The foliation tends to follow the contact on a regional scale (Pl. 1). In the northern half of the area the foliation strikes more northerly than in the southern half, and the main contact also trends more northerly in the northern half.

Many small discordant contacts are recognized, but the ratio of the lengths of concordant contacts to discordant contacts is very large. The angle of discordancy is small and is generally seen only upon close examination (Fig. 23). The lobed margin of the main contact of the complex north and east of Alexander City is similar to that shown in Figure 23b but on a much smaller scale.

The eastern contact of the complex in northern Elmore County (Pl. 1) seems to be slightly discordant. The contact is poorly exposed, and it is uncertain whether this

is a true discordant contact or is due to many small local variations in the attitude of the schist foliation.

To summarize, all contacts both external and internal are essentially concordant and conformable, but all combinations of concordant, discordant, conformable, and disconformable occur throughout the complex.

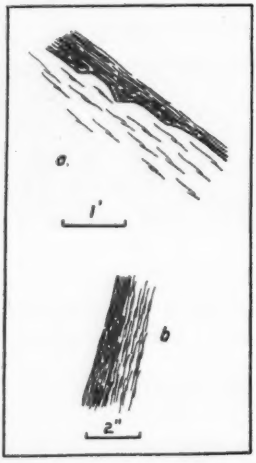


FIGURE 23.—*Typical discordant contacts*
a. Sec. 16, T. 23 N., R. 20 E., Coosa County.
b. Elkahatchee Creek, Tallapoosa County.

At many places the main contact is inferred because of insufficient exposures and because of alternations of schist and igneous rocks. Along the eastern margin the boundary is usually distinct, but in some places schist septa make it difficult to locate the boundary accurately particularly where exposures are scarce. The western boundary is well defined from Clay County to Goodwater. South from Goodwater there are many intercalations of gneiss and schist, and the boundary is a zone rather than a line. This intercalated pattern continues south to the Coosa River (Pl. 1). Along the river on the west side of the complex, at least 10 schist septa 100 feet or more wide are recognized within a distance of a mile across the strike. The minimum width of these schist septa along the river is about 100 feet.

No detailed work has been done west of the area shown on Plate 1. A few observations indicate that the intercalation of gneiss and schist continues west from Rockford for several miles. The gneiss bands become less numerous and narrower and are granodiorites and granites rather than quartz diorite.

PETROFABRIC ANALYSIS

GENERAL CONSIDERATIONS

Within the last decade the application of petrofabric analysis has become increasingly important as an aid to the interpretation of geologic structures. To supple-

ment the field observations, fabric analyses were made for 15 oriented specimens of the Pinckneyville complex and 4 oriented specimens of the wall rocks. The specimens were chosen as typical petrographic types and as showing representative structures. The localities at which the specimens were collected are distributed throughout the area so that the fabric analyses give an overall picture of the microscopic fabric of the area. The theory and method of petrofabric analysis have been discussed fully by other investigators (Sander, 1930; Knopf, 1933; Fairbairn, 1935; 1937; Knopf and Ingerson, 1938; and others).

All measurements were determined with a four-axis universal stage and were plotted on an equal-area net representing the lower hemisphere. The concentrations within a unit area were counted at half-centimeter intervals, and the contours are drawn on equal per cents of concentration. The mica diagrams represent the poles of cleavage flakes, and the quartz diagrams show the distribution of the optic axes. All diagrams were made from thin sections cut parallel to the *ac*-plane as determined in the field and are normal to the foliation and lineation. The regional *b* axis lies approximately in the center of each diagram.

The geographic orientation of each diagram is shown by an arrow and dip line in the center of the diagram. The arrow always points east; thus, 78 is N. 78° E. and 114 is S. 66° E. All diagrams are viewed from the southwest or southeast. Dip angles up to 90° indicate a plane dipping toward the reader. Angles greater than 90° represent overhanging planes. Each diagram is numbered, and, referring to diagrams in the text, the number is preceded by the letter D.

MICA ORIENTATIONS

General discussion.—Biotite and muscovite are analyzed separately for 16 of the 19 specimens. In the other three specimens only biotite was analyzed. The mica diagrams are all *ac*-girdles of varying degrees. (Statistical concentrations of points around a great circle of the projection normal to *b* give rise to an *ac*-girdle; statistical concentrations of points around a point give rise to a point maximum). The maximum girdle widths of the 1, 2, and 4 per cent contours measured on the periphery are shown in Figures 24–26. D 46 is the only diagram with a 1 per cent girdle of less than 180°, and there are only four diagrams (D 25, D 46, D 54, and D 56) with 1 per cent girdles of less than 240°. Three diagrams (D 9, D 13, and D 34) show complete 360-degree girdles.

In general, the girdle widths tend to decrease as the maxima values increase, but several diagrams are at variance with this tendency. D 7 shows the lowest mica maximum with the smallest 4 per cent girdle. D 12 and D 13 have high maxima (17 and 14 per cent) with 1 per cent girdles greater than 300°. Their 2 and 4 per cent girdles are among the largest in the mica diagrams.

The total area included within 4 per cent and higher contours is not confined to one area, except in D 7. The total area is made up of several small areas which are scattered around the periphery and are usually more abundant near the major maxima (D 33, D 34, D 37, and D 42). The foliation in the hand specimens is indistinct.

Biotite girdles tend to be more complete than muscovite (Table 2), but there is

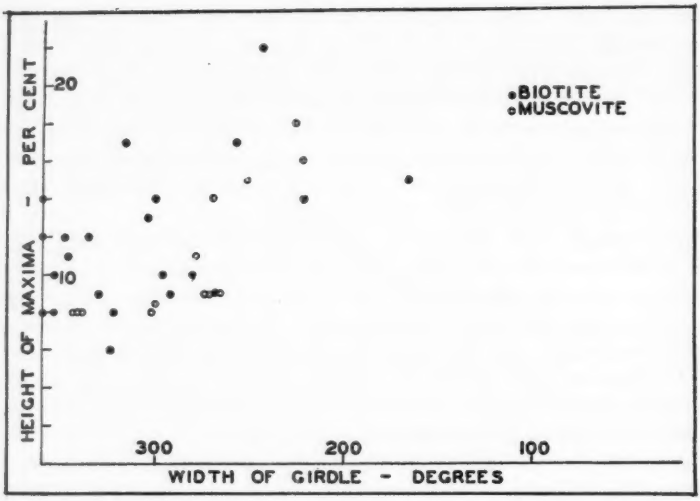


FIGURE 24.—*Width of 1 per cent mica girdles*

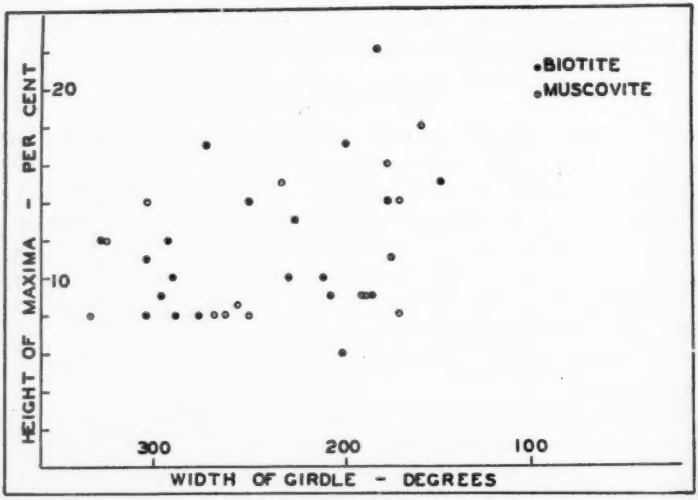


FIGURE 25.—*Width of 2 per cent mica girdles*

only a few degrees difference in width between complementary muscovite and biotite diagrams. Many biotite maxima are higher than muscovite, but some muscovite maxima equal biotite or are higher (Tables 2, 3).

The number of major maxima varies from one to four in biotite and muscovite diagrams, but over half the diagrams have only one major maximum. More than one major maximum in a biotite diagram does not signify that the complementary

muscovite diagram will have more than one. Diagrams with high values for the major maximum usually show only one.

D 53 and D 54, and D 56 and D 57, are the only complementary pairs of mica diagrams in which the maxima exactly coincide upon superposition of the diagrams. A

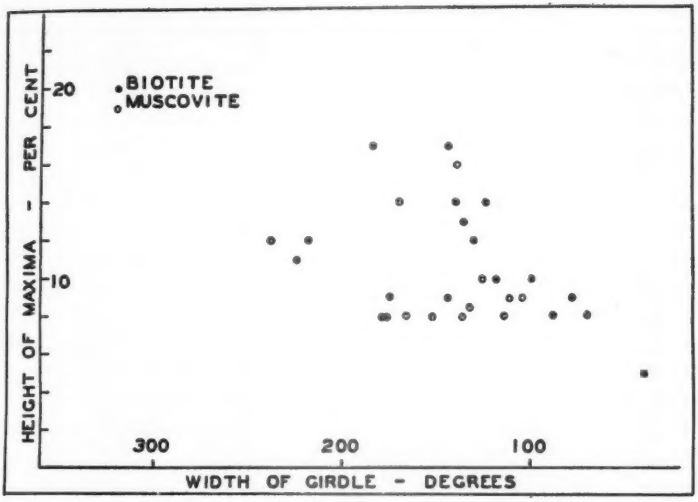


FIGURE 26.—*Width of 4 per cent mica girdles*

TABLE 2.—*Comparison of relative girdle widths*

	Girdles				
	1 per cent	2 per cent	4 per cent	6 per cent	Major maximum
Muscovite greater than biotite	7	5	4	1	3
Biotite greater than muscovite	9	11	10	1	8
Biotite equals muscovite	—	—	—	—	5

rotation of 10°–90° around *b* of one diagram with respect to its complementary diagram is necessary in the remainder to bring the maxima into coincidence. It is necessary in some cases to rotate on an axis normal to *b* as well as to rotate on *b*. Rotations of 10°–80° on *b* and other axes are necessary to bring the areas included within the 2 and 4 per cent contours into general coincidence. Areas within equal contours do not always coincide after the maxima are rotated into coincidence. The direction of rotation for biotite with respect to muscovite is counterclockwise for some pairs of diagrams and clockwise for others. The direction of rotation shows

no relationship to the attitude of the *s*-planes determined by the maxima or to the *s*-plane determined by the foliation observed in the field. (An *s*-plane is the plane represented by the concentration of mica points in a great circle of the projection 90° from the maximum; other planar directions can be considered *s*-planes such as joints,

TABLE 3.—*Height of maxima and number of occurrences*

Maximum value	Muscovite	Biotite	Quartz
(*per cent*)			
3½	—	—	1
4	—	—	5
4½	—	—	5
5	—	—	4
5½	—	—	2
6	—	1	4
8	5	2	—
8½	1	—	—
9	2	3	—
10	1	3	—
11	1	1	—
12	1	2	—
13	—	1	—
14	2	2	—
15	1	1	—
16	1	—	—
17	—	2	—
18	1	—	—
22	—	1	—

bedding, and cleavage.) The relationships are further complicated where more than one major maximum occurs in complementary diagrams. Maxima are located at the clockwise end, in the center, or at the counterclockwise end of the areas included within the 4 per cent contour.

There is no correlation between the attitude of the *s*-planes determined in the diagrams with the geographic distribution of the specimens, the structures within the hand specimens, or between complementary mica diagrams. The attitudes of many *s*-planes determined by the biotite maxima differ from those determined by the maxima in the complementary muscovite diagram.

The girdle-type pattern of the mica diagrams is distributed throughout the area. There is a suggestion that the girdles are more complete toward the northwest and northeast. One per cent girdles show this distribution better than 2 or 4 per cent girdles. No other geographic relationship is apparent.

The behavior of a mineral under a given set of conditions is greatly influenced by the associated minerals and rock type (Sander, 1930; Cloos and Hietanen, 1941). The mineral assemblage of quartz, feldspar, and mica and the origin of any rock of the Pinckneyville complex are inherently the same as any other rock of the complex. The schist specimens also contain about the same minerals as the Pinckneyville rocks. This similarity of the mineralogy and origin of the igneous rocks and of the mineralogy

of the metamorphic rocks and igneous rocks gives a common ground on which to compare the diagrams.

D 59 is a composite diagram (Knopf and Ingerson, 1938, part II) prepared by rotating all major mica maxima into a horizontal projection. The diagram shows clearly the *ac*-girdle which dominates the fabric of the Pinckneyville complex. The attitudes of *s*-planes determined by muscovite points in the composite diagram are slightly steeper than the biotite *s*-planes.

In summary, the biotite and muscovite fabrics of the Pinckneyville complex are *ac*-girdles. There is no constant relation between the biotite and muscovite orientations with respect to the amount or the direction of rotation for coincidence of their maxima, the geographic distribution of girdles, or the structural features of the specimens they represent.

Where the foliation of the gneiss is indistinct and poorly developed, the fabric analyses show well-developed girdle patterns. In rocks with a megascopically recognizable foliation, the girdles are smaller and the maxima tend to be higher. The lineation, recognizable in many exposures, is verified by the girdle pattern as it indicates an arrangement of the micas around an axis. This microscopic axis approximately parallels the megascopic *b* axis determined in the field.

There is a slight difference between the strikes of the microscopically determined *s*-planes and the strikes of foliation recorded in the field. The strikes of the *s*-planes determined by the mica maxima are strewn more widely than those of the foliation determined in the field. This is due in part to the small area over which the measurements are made in each thin section compared with the larger area over which the measurements in the outcrop were made. The measured megascopic foliation represents the trend of the foliation for several feet along which there are minor fluctuations in the orientations of the mica aggregates due to their bending around feldspar crystals or quartz patches. The thickness of a thin section cut across the strike represents a distance of only a few microns along the strike of the rock structures. In a diagram prepared for such a thin section the mica may deviate from the average strike, and there will then be an angular difference between the microscopic *s*-plane or lineation and that determined in the field.

Some of the discrepancies between the attitudes of the megascopic foliation and the attitude of the microscopic *s*-planes are due to the presence of several maxima in one diagram. The foliation in such cases is determined for the most prominent *s*-plane or is an average of all the *s*-planes. In a few cases, the megascopic foliation corresponds to the *s*-plane determined by biotite because it is more easily recognized than muscovite. Where complementary mica diagrams each contain several maxima and they do not coincide upon superposition, the rock which they represent has a faint foliation or none at all, but the lineation is well developed (D 33, D 34).

Descriptions of the individual mica diagrams are given in Table 4.

Interpretation of the mica fabric.—The mica fabric of the Pinckneyville complex is that of an *ac*-girdle. Girdle fabrics may develop as cylindrical flow or as rotations, and their interpretation must be correlated with structural field observations. Mica flakes oriented in a stream will align themselves with their longest dimensions parallel to and their shortest dimensions normal to the direction of flow. Because

TABLE 4.—*Description of mica diagrams by locality*

Locality	Rock type	Diagams	Contour Width of girdles on periphery in degrees			Maxima per cent	Remarks
			1	2	4		
19 sec. 5, T. 19 N., R. 18 E., Elmore County	quartz diorite gneiss	D 1	296	210	100	10	Foliation distinct, biotite maximum rotated counterclockwise from muscovite
		D 2	272	186	110	9, 8, 8.	
34 sec. 16, T. 20 N., R. 18 E., Elmore County	quartz diorite gneiss		1	2	4		Foliation visible, biotite maxima rotated clockwise from muscovite
		D 4	268	206	144	9, 8, 8, 8.	
		D 5	274	186	104	9, 8, 8, 8.	
11 sec. 24, T. 20 N., R. 18 E. Elmore County	hornblende-biotite gneiss		1	2	4		Foliation indistinct, lineation good.
		D 7	324	200	50	6, 6.	
46 sec. 9, T. 20 N., R. 17 E. Elmore County	quartz diorite gneiss		1	2	4		Biotite maximum rotated clockwise from muscovite; foliation indistinct.
		D 9	354	288	88	8	
		D 10	280	228	126	10	
31 sec. 14, T. 20 N., R. 17 E., Elmore County	granite dike		1	2	4		Biotite maximum rotated counterclockwise from muscovite, lineation distinct, foliation weak. Dike vertical and strikes 17°, dips 45° NW. Average attitude of mica *s*-planes strikes 36°, dips 28° NW.
		D 12	316	272	184	17	
		D 13	360	304	170	14	
33 sec. 14, T. 20 N., R. 17 E., Elmore County	gray quartz diorite inclusion		1	2	4		Muscovite lacking; many bent and faulted biotites. D 15 shows 74 undeformed biotites; D 16 shows the 74 points plus 54 points measured on bent or faulted segments of 22 deformed biotites; segments also show girdle arrangement; inclusion spindle-shaped rather than disc-shaped; foliation indistinct. D 15 not represented on Figures 24, 25, and 26, or on D 59
		D 15	334	284	222	14	
		D 16	348	328	218	12	
47 sec. 6 T. 20 N., R. 18 E., Elmore County	fine-grained quartz diorite		1	2	4		Biotite maximum rotated counterclockwise from muscovite.
		D 18	336	292	130	12	
		D 19	342	268	166	8, 8.	

R. 18 E., Elmore County	diorite	D 18	336	292	130	12	
		D 19	342	268	166	8, 8.	
48 sec. 8 T. 20 N., R. 19 E., Elmore, County	fine-grained quartz diorite		1	2	4		Biotite maximum rotated counterclockwise from muscovite, foliation distinct.
		D 21	304	226	136	13	
		D 22	300	256	132	8.5	
39 sec. 14 T. 20 N., R. 18 E., Elmore County	fine-grained light-gray gneiss		1	2	4		Biotite maximum rotated clockwise from muscovite, principal area within biotite 4 per cent contour counterclockwise from muscovite; foliation quite distinct; locality 39, 20 feet from locality 17.
		D 24	258	198	144	17	
		D 25	222	176	140	16, 16.	
20 sec. 11 T. 20 N., R. 18 E. Elmore County	dark quartz diorite gneiss		1	2	4		Biotite maximum rotated clockwise from muscovite; foliation distinct (*see* locality 39)
		D 27	354	290	118	10	
		D 28	278	174	90	11	
40 sec. 18 T. 22 N., R. 20 E. Coosa County	quartz diorite gneiss		1	2	4		Biotite maximum rotated clockwise from muscovite; foliation indistinct.
		D 30	360	304	140	8	
		D 31	342	262	136	8	
30 sec. 17, T. 22 N., R. 20 E., Coosa County	fine-grained granite gneiss almost alaskite		1	2	4		Muscovite maximum rotated 35° counterclockwise from nearest biotite; strong lineation, no foliation.
		D 33	346	304	224	11, 11.	
		D 34	360	326	238	12	
15 sec. 12 T. 21 N., R. 20 E., Coosa County	quartz diorite gneiss		1	2	4		Muscovite maximum rotated 20° clockwise from nearest biotite; foliation visible.
		D 37	322	272	178	8, 8, 8.	
		D 38	302	250	114	8	
26 Rockford, Coosa County	granodiorite gneiss		1	2	4		Biotite maxima rotated 20° to 40° counterclockwise from muscovite; foliation suggested by biotite. Westernmost sample with respect to Pinckneyville complex.
		D 41	330	296	174	9, 8, 8.	
		D 42	340	334	152	8, 8.	
41 sec. 6 T. 21 N., R. 19 E., Coosa County	hornblende-biotite gneiss		1	2	4		Lineation well developed, foliation indistinct.
		D 45	300	250	140	14, 13.	

TABLE 4—*Continued*

Locality	Rock type	Diagrams	Contour Width of girdles on periphery in degrees			Maxima per cent	Remarks
			1	2	6		
42 sec. 7 T. 22 S., R. 7 E., Clay County	quartz-mica schist muscovite porphyroblasts	D 46	166	148	104	15, 15.	Biotite maxima rotated a few degrees clockwise from muscovite. D 48 shows 38 muscovite porphyroblasts superposed on diagram showing 1 per cent contour and maxima of D 7; porphyroblasts strained, show a slight tendency to produce a girdle. Locality 42 is a third of a mile west of western margin.
		D 47	252	232	92	15, 14, 14.	
			1	2	6		
43 Elkahatchee Creek, Tallapoosa County	quartz-garnet mica schist	D 53	244	182	78	22	Bedding and flow cleavage parallel, no fracture cleavage, maxima coincide; 200 feet east of contact; *see* locality 44.
		D 54	226	158	98	18	
			1	2	4		
44 Elkahatchee Creek, Tallapoosa County	granodiorite border gneiss	D 56	222	176	124	14	Locality 44 is 240 feet from contact. Maxima coincide, area within the biotite 4 per cent contour rotated 15° clockwise from muscovite, foliation strong.
		D 57	270	170	116	14, 14.	
			1	2	4		
45 sec. 4 T. 22 S., R. 7 E., Clay County	quartz diorite gneiss	D 50	292	184	78	9	At extreme northeastern end of complex, about 100 feet from schist wall rock. Biotite maximum rotated 40° clockwise from muscovite. Foliation indistinct.
		D 51	266	170	70	8	

the cleavage pole and the shortest dimension are the same, this arrangement of mica flakes results in a girdle of poles normal to the direction of flow. Flow structures are usually best near the margins of many igneous bodies because of more intense friction. Magmatic flow structures are recognized in the Pinckneyville. However, they are not the conspicuous ones but are masked by later structures. Furthermore, the more complete girdles are in general located away from the margins of the complex, and the less complete girdles are developed along the margins. The *b*-axis is horizontal or nearly horizontal, and such a direction of magmatic flow does not seem likely, particularly because it is uniform throughout an area of several hundred square miles. Thus the girdle fabric is probably not primarily the result of orientation in the flowage phase but in a later phase.

It is also difficult to attribute the development of the Pinckneyville fabric to rotation by deformation—*i.e.*, regional deformation of a solid mass. Complete girdle patterns are described by other workers as representing an intense deformation (Cloos and Hietanen, 1941) and are considered rotation fabrics due to deformation (Sander, 1930; Knopf and Ingerson, 1938; Turner, 1940; Cloos and Hietanen, 1941). In the Pinckneyville the rocks showing the more intense deformation show planar fabrics (border gneisses and metamorphic rocks). The areas described by Cloos and Turner are metamorphic rock terranes, and the rotations of the minerals are recognized by a definite sequence of crystallization (recrystallization or mechanical deformation of the minerals). Mechanical deformation of the micas in the Pinckneyville is slight and had little effect on their orientation (D 15 and D 16), there is no well-defined sequence of crystallization in comparison with those of Cloos and Turner, and there is no suggestion of recrystallization.

The parallelism of structures in the gneiss with those in the schist and the mutual relations between gneiss and schist suggest that regional deformation was in part effective in developing the structures of the gneiss. The mica orientations which give the girdle fabric are considered to have developed in the transition phase. Biotite and muscovite maxima do not coincide upon superposition of complementary pairs of diagrams. In a partially consolidated magma subjected to regional deformation, the igneous material would yield by flowage rather than by shearing and recrystallization. The direction of flow would be the direction of easiest relief, which in this case is nearly horizontal and is parallel to *b*. This direction is also indicated by other structures such as boudins of inclusions. The micas were oriented with their long dimensions parallel to the direction of flow and the short dimensions in a girdle normal to it, and the structures produced are both metamorphic and igneous.

QUARTZ ORIENTATIONS

General discussion.—Fabric studies are not so applicable to quartz as to mica orientations. Quartz lacks good cleavage. It fractures less regularly than mica, and the fractures can follow several crystallographic directions. This larger number of probable directions results in more possible orientations of quartz and thus hinders concentrations of points in well-defined maxima. Quartz deformation and orientations, as well as mica, are influenced by the associated minerals. These are the same throughout the area so that all fabric diagrams have a similar petrographic background. Differences or similarities in the diagrams can therefore probably be inter-

preted as due to factors other than mineral environment. The quartz occurs in *ac*-girdles which are less complete girdles than the mica girdles. The composite diagram, D 60, represents the horizontal projection of all quartz maxima of 4 per cent or higher and three 3½ per cent maxima from D 49 and is an *ac*-girdle. It is

TABLE 5.—*Position and heights of quartz maxima*
Maxima according to Fairbairn (1939)

Diagram	Locality	Percentage						
		I	II	III	IV	V	VI	Unclassified
D 3	19	6	4					
D 6	34		4		5			
D 8	11				4½			4½
D 11	31	4½	4½		3		4½	
D 14	31		4		4		4	
D 17	33		3	3			4½	
D 20	47	6½			4			
D 23	48	5½	5½		5½			
D 26	39	5	4			3	3	
D 29	20		3		4			
D 32	40	3			5	3		4
D 35	30		5		4		4	
D 36	30				5		5	5
D 39	15	4	4					
D 40	15		4			4	5	4
D 43	26		4	6	5		4	5
D 44	26	4		5	4			
D 49	42				3½			
D 52	45				4	3		
D 55	43		4		4			
D 58	44	4½	4					
Totals.....		9	14	3	15	4	8	5

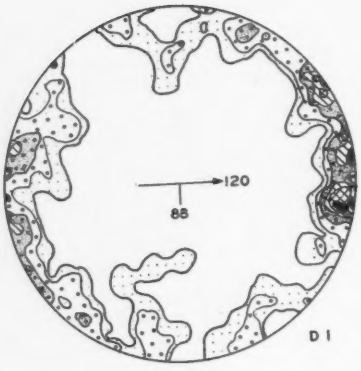
less complete than the composite mica diagram, D 59. The points spread toward the periphery corresponding to the spreading of points in the composite mica diagram, but *b* is empty. The similarity of the two composite diagrams suggests a relationship between the orientation of the mica cleavages and the orientation of the quartz fragments.

Griggs and Bell (1938) and Fairbairn (1939) have predicted the locations of quartz maxima on the assumption that quartz is oriented by deformation in crystallographic directions, as for instance the possible crystal faces *c*, *a*, *m*, *z*, *r*, etc. The maxima in the quartz diagrams are classified according to this hypothesis in Table 5, and it is assumed that *a* is the direction of transport (Fairbairn, 1939, Fig. 4). There is only one position for each quartz diagram in which the measured maxima coincide closely with the predicted locations of Fairbairn, and the *a* of the measured maxima was determined from this position for each diagram.

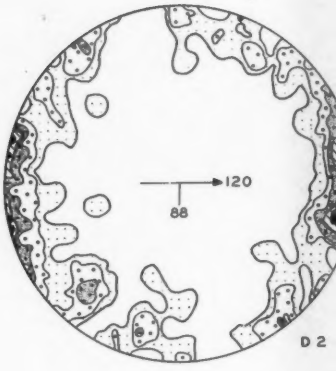
The statistically determined maxima coincide exactly only for a few diagrams. (*Cf.* D 55.) About half the classified maxima fall near the periphery. Comparison

Diagram number	Number and type of grains measured	Contours in per cent	Rock* type	Locality
D 1	100 biotite	1–2–4–6–8–10	qd.	sec. 5, T. 19 N., R. 18 E.
D 2	100 muscovite	1–2–4–6–8	qd.	Elmore County
D 3	200 quartz	1–2–3–4–5–6	qd.	
D 4	100 biotite	1–2–4–6–8	qd.	Sec. 16, T. 20 N., R. 18 E.
D 5	100 muscovite	1–2–4–6–8	qd.	Elmore County
D 6	222 quartz	1–2–3–4–5	qd.	
D 7	150 biotite	1–2–4–6	hbn.	
D 8	200 quartz	1–2–3–4	hbn.	Sec. 24, T. 20 N., R. 18 E. Elmore County
D 9	100 biotite	1–2–4–6–8	qd.	Sec. 9, T. 20 N., R. 17 E. Elmore County
D 10	100 muscovite	1–2–4–6–8	qd.	
D 11	200 quartz	1–2–3–4	qd.	
D 12	100 biotite	1–2–4–8–12–16	qd.	
D 13	100 muscovite	1–2–4–8–12	qd.	Sec. 14, T. 20 N., R. 17 E. Elmore County
D 14	200 quartz	1–2–3–4	qd.	
D 15	74 biotite	1–2–4–8–12–14	inc.	
D 16	128 biotite	1–2–4–8–12	inc.	Sec. 14, T. 20 N., R. 17 E. Elmore County
D 17	200 quartz	1–2–3–4	inc.	
D 18	100 biotite	1–2–4–8–12	qd.	
D 19	100 muscovite	1–2–4–6–8	qd.	Sec. 6, T. 20 N., R. 18 E. Elmore County
D 20	203 quartz	1–2–3–4–5–6	qd.	
D 21	100 biotite	1–2–4–6–8–12	qd.	
D 22	100 muscovite	1–2–4–6–8	qd.	Sec. 8, T. 20 N., R. 17 E. Elmore County
D 23	200 quartz	1–2–3–4–5	qd.	
D 24	100 biotite	1–2–4–8–12–16	gg.	Sec. 11, T. 20 N., R. 18 E. Elmore County

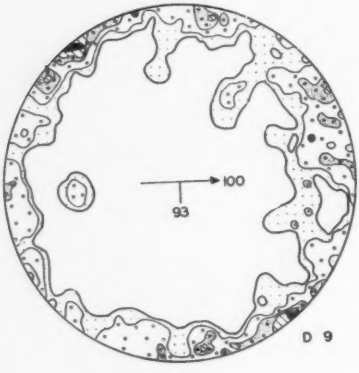
* qd.—quartz diorite; hbn.—hornblende gneiss; inc.—inclusion; gg.—gray gneiss.



D 1



D 2



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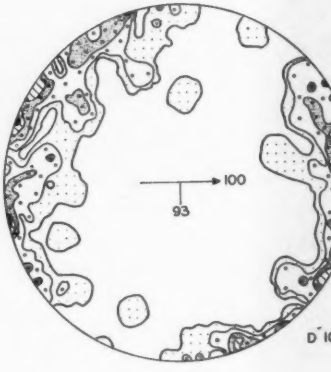

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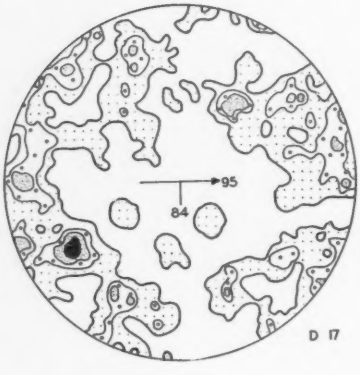

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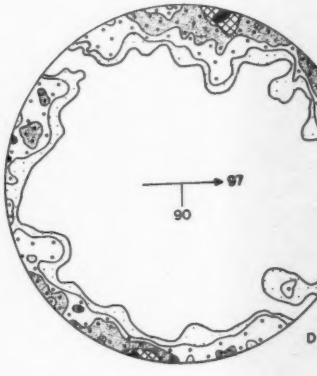


D

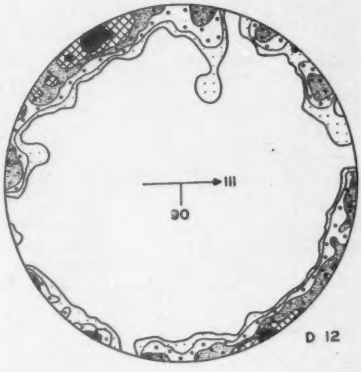
D 2

D 3

D 4

D 10

D 11



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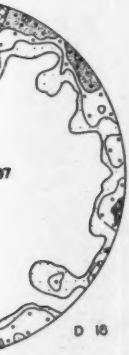


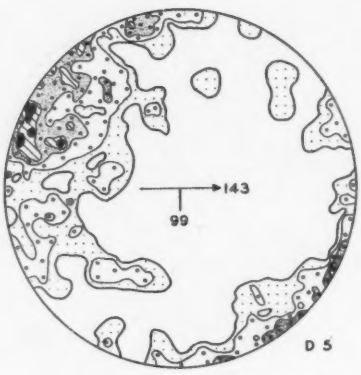
D 18

D 19

D 20

D 3

D 4



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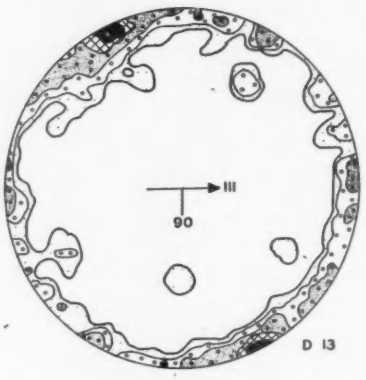
D 11

D 12



D 13

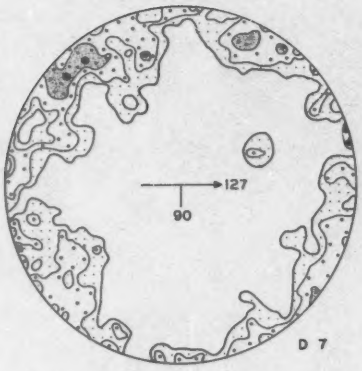
D 19

D 20

D 21

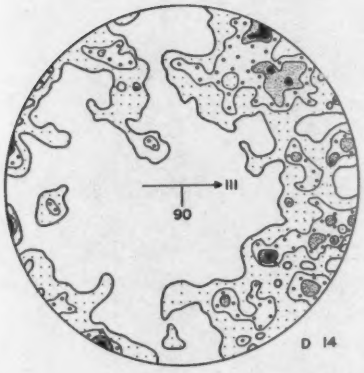
FABRIC DIAGRAMS

D 6



D 7

D 8



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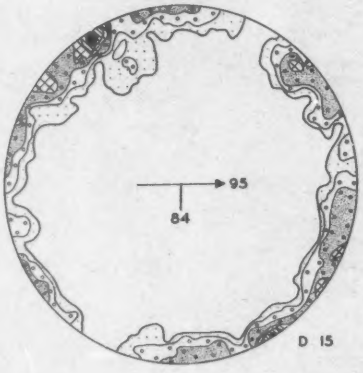

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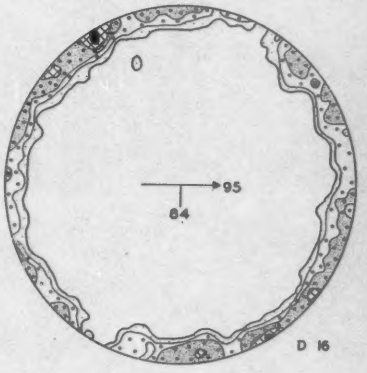

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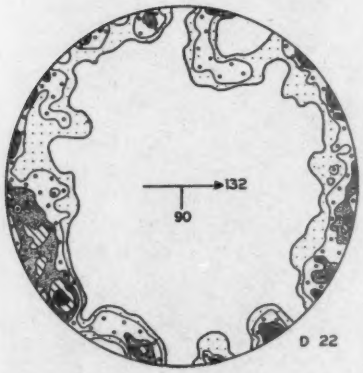
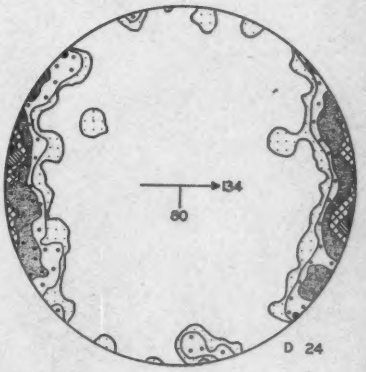


D 22

D 23



D 24

of the position of *a* determined for the measured maxima with the attitudes of known *s*-planes shows four possibilities: (1) in the plane determined by the megascopic foliation, (2) in an *s*-plane determined by a biotite maximum, (3) in an *s*-plane deter-

TABLE 6.—*Position of inferred direction of movement for quartz and attitude of mica s-planes and foliation*

Diagram	a	S	bi-s	mu-s
D 3	60 SE.	vertical	vertical to steep NW.	vertical to steep NW.
D 6	60 NW.	35 NW.	60 NW.	steep NW.
D 8	horizontal	45 NW.		
D 11	40 NW.	80 SE.	40 NW.	(1) steep NW. (2) steep SE.
D 14	80 SE.	45 NW.	30 NW.	25 NE.
D 17	20 SE.	40 NW.		
D 20	8 SE.	30 SE.	20 SE.	(1) 30 SE. (2) 50 SE.
D 23	40 SE.	40 SE.	40 SE.	40 SE.
D 26	65 SE.	70 NW.	75 NW.	(1) 85 NW. (2) vertical
D 29	50 SE.	75 NW.	(1) 65 NW. (2) 75 NW.	
D 32	35 SE.	60 SE.	50 SE.	35 SE.
D 35	30 NW.		(1) 50 SE. (2) 65 NW.	75 NW.
D 40	45 SE.	75 SE.	(1) 30 SE. (2) 40 SE. (3) 70 SE.	vertical
D 43	85 SE.	65 SE.	(1) 5 SE. (2) 40 SE. (3) 55 SE.	(1) 60 SE. (2) 80 NW.
D 49	20 SE.	40 SE.	50 SE.	(1) 15 SE. (2) 25 SE. (3) 45 SE.
D 52	60 SE.	50 SE.	60 SE.	40 SE.
D 55	40 SE.	50 SE.	50 SE.	50 SE.
D 58	45 SE.	45 SE.	45 SE.	45 SE.

mined by a muscovite maximum, or (4) in a position unrelated to any known *s*-plane. There is no constant direction or angular relationship between *a* and the attitude of *s*-planes determined by the foliation or mica maxima. Table 6 shows the position of *a* according to the classification in Table 5 with respect to the foliation and mica *s*-planes. In Table 6, *a* is the direction of movement; *S* is the observed foliation; *bi-s* is an *s*-plane determined by a biotite maximum; *mu-s* is an *s*-plane determined by a muscovite maximum; parenthetical numbers are used where there is more than one mica maximum. (1) refers to the maximum nearest the horizontal line of the diagram, (2) is the next closest, etc.; 60 SE is the dip of *a* below the horizontal measured on the periphery of an equal-area net in the quadrant in which the measurement was made.

D 36, D 39, and D 44 are selective diagrams prepared for quartz. Many of the

axes of individual grains within large patches or within portions of large patches fall close to one another. These grains appear to be distinct individuals rather than parts of a strained crystal, but they represent the fragments of a larger grain which has been fractured.

The selective diagrams were prepared from each corresponding nonselective diagram by eliminating all but one point from any group of points representing a large patch of quartz fragments in which more than one point lies within a circle of radius one tenth the radius of the equal-area net. The point retained represents an average. Each large quartz patch was treated in the same manner. The points remaining, including those representing small isolated grains, were counted and contoured as in the other diagrams.

There is little or no difference between the outline of the 1 per cent contour in the selective diagram and the nonselective diagram at each locality. The location of the major and minor maxima does not change, but the heights of the maxima do. Nine maxima in the three selective diagrams are lower than the corresponding maxima in the nonselective diagrams, two show no change, and the heights of four maxima increase in the selective diagrams. Where the per cent of loss of points in the maxima in the selective diagram is greater than the per cent of loss of the total number of points, the maxima decrease in height and vice versa.

Interpretation of quartz diagrams.—Interpretation of the quartz diagrams is difficult, and reference to the mica diagrams and structural field observations is utterly essential. The strong undulatory extinction of quartz indicates that it was affected by stress and that recrystallization was slight or incomplete (Eskola, 1915; Fairbairn, 1935). Observations of others (Sander, 1930; Sahama, 1936; Hietanen, 1938; Griggs and Bell, 1938) suggest that the fair degree of correspondence between measured and predicted quartz orientation permits the application of the fracture hypothesis.

The three selective quartz diagrams indicate that some large quartz grains were fractured but that the fragments were rotated only a few degrees. Several diagrams show a trend from the maxima IV and VI zones (Fairbairn, 1939, Fig. 4) to maxima locations on the periphery. The trend of maxima moving toward the periphery and the evidence of the selective diagrams suggest that quartz orientations were not complete. The nonselective diagrams, however, indicate that deformation was sufficient to develop some orientation of the quartz.

The quartz orientations may be related to *s*-planes determined statistically or in the field. Fractured grains are thought to rotate into shearing planes with their long axes parallel to the direction of movement or *a* (Sander, 1930; Fairbairn, 1939). The mica *s*-planes and the foliation are generally considered to represent shear planes in which the micas grew.

It has been shown (Table 6) that the quartz orientations are not always related to known *s*-planes. Quartz was one of the last minerals to cease crystallization, if not the last. Most of the orientation of mica took place in the transition phase. Quartz, however, could not be deformed and oriented until it had crystallized or until the rocks had about solidified. Thus the quartz orientations are probably later than the mica orientations and agree with mica *s*-planes only where the direction of movement

due to deformation fell into an *s*-plane or where a mica *s*-plane was prominent enough to influence the direction of transport. Other late developments of quartz orientations are described by Turner (1940).

SUMMARY AND CONCLUSIONS

The Pinckneyville quartz diorite complex intrudes a series of mica, quartz, graphite, and garnet schists and phyllites and smaller amounts of hornblende gneisses and other metamorphic rocks. Schist septa varying considerably in size occur within the complex. The rocks of the septa are similar to those on the margins.

The principal minerals of the Pinckneyville rocks are andesine, oligoclase, microcline, quartz, biotite, muscovite, and small but persistent amounts of epidote. Accessory minerals are allanite, sphene, zircon, apatite, chlorite, pyrite, garnet, and tourmaline. A dark-gray, coarse-grained quartz diorite composed of plagioclase, quartz, biotite, and muscovite constitutes most of the complex. Granodiorites and granites occur in smaller amounts and are lighter-colored, containing plagioclase, orthoclase, microcline, muscovite, quartz, and biotite. The textures and structures vary from subporphyritic to strongly gneissic with well-developed augen. All textures and structures occur in all normal rock types and in many of the dikes.

The hornblende gneisses are subdivided into three types: (1) amphibolites, (2) hornblende-epidote gneiss, and (3) hornblende-biotite gneiss. The occurrence of olivine and pyroxene and the asscciation of cortlandite with hornblende gneisses near by suggest that they are probably metamorphosed basic sills and dikes which preceded the quartz diorite.

Granodiorite, granite, aplite, pegmatite dikes, and small quartz veins are abundant throughout the area. Schist, hornblende gneiss, and quartz diorite inclusions occur within the rocks of the complex. Minor amounts of injection gneiss, mylonites, and hydrothermally altered basic rocks are also present.

The sequence of intrusion from oldest to youngest is (1) quartz diorite, (2) granodiorite, (3) granite. These stages overlap, and each was accompanied by pegmatite and aplite intrusion. Northeast trends dominate the orientation of planar structures, lineations, inclusions, and contacts. Flow cleavage of the wall rocks usually strikes northeast and dips steeply southeast parallel to the contacts. Fracture cleavage was superposed on the flow cleavage.

Planar flow structures within the complex are shown in the orientation of inclusions and micas. Lineation due to the arrangement of micas around *b* is prominent and developed principally in the transition phase. Border gneisses also developed in the transition phase along the margins and locally within the complex. They represent zones of yielding to regional forces.

Joints in the schist and in the gneiss are widely distributed and can be grouped into definite sets: cross joints, oblique joints, and longitudinal joints of several types related to other rock structures. Pegmatites and other dikes follow the joint directions.

All structures occur throughout the area, but the intensity of foliation decreases from east to west. The foliation strikes northeast and dips steeply southeast except in Elmore County where northwest dips are observed. Structures within schist

inclusions indicate that folding commenced before the intrusion of the Pinckneyville quartz diorite.

Internal and external contacts are essentially concordant and conformable. Locally discordant and disconformable contacts show slight angular cross-cutting relations and are additional evidence for an intrusive origin of the Pinckneyville complex. The eastern boundary of the complex is usually well defined. The western boundary is a zone containing many septa of schist intercalated with gneiss.

The Pinckneyville quartz diorite is characterized by both igneous and metamorphic features, indicating that it is a syntectonic intrusion—*i.e.*, it was intruded during the folding and metamorphism of the region (Högbom, 1928; synchronous, Billings, 1928; H. Cloos, 1936; syntectonic, Williams and Billings, 1938). The structural features are thought to be the combined result of magmatic flow and contemporaneous folding.

Several facts point to the intrusive origin of the rocks. The many discordant contacts are intrusive contacts. The great aggregate length of the concordant and conformable contacts compared with the discordant contacts is the result of intrusion into a series of rocks which had been folded and were still being folded. The inclusions also indicate an intrusive origin, and the different types of inclusions show the sequence of intrusion. Hornblende gneiss and schist inclusions occur in quartz diorite, and inclusions of the quartz diorite occur in the granodiorite. Schist inclusions also are found in the granodiorite.

The sharp petrographic boundaries between wall rocks and rocks of the complex are intrusive. Absence of contact zones around the inclusions and reaction zones within the inclusions suggest the inclusions were picked up mechanically and were in equilibrium with the magma.

The normal zoning of the feldspars is typical of igneous rocks. Decrease in the amount of plagioclase and biotite and increase in the biotite indices indicate the same sequence as do the inclusions. This sequence is in accord with the generally accepted ideas of magmatic differentiation. The uniform composition of the quartz diorite throughout the area denotes a common source.

Some of the structures of the Pinckneyville quartz diorite are igneous flow structures, while others are due to regional deformation. Planar flow structures occur throughout the area, but in general the Pinckneyville complex is only slightly foliated. Well-developed platy flow structures, however, are not to be expected where the walls of the intrusive are soft and can yield readily to the expanding magma (Balk, 1937). The complex does contain a foliation and a prominent lineation observed in the field and in the fabric diagrams as later than and masking the flow structures. The orientation of the micas which produces these structures apparently took place before solidification. Orientation of the inclusions cannot be explained by rotation in a solid body, and it is doubtful if small flexures and drag folds of the foliation and lineation developed after the rock was solid. The boundinage structures imply flowage of the quartz diorite and stretching of the inclusions parallel to *b*. These structures are most readily explained as developing during the transition phase.

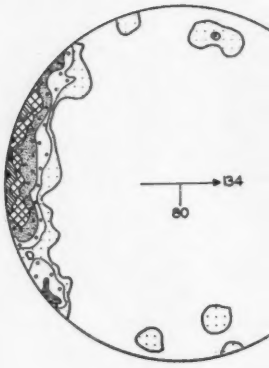
Many of the border gneisses are near contacts with the wall rocks. The greatest interference of crystals with one another in a mush or the greatest resistance to a

Diagram number	Number and type of grains measuured	Contours in per cent	Rock* type
D 25	100 muscovite	1–2–4–8–12–16	gg.
D 26	212 quartz	1–2–3–4–5	gg.
D 27	100 biotite	1–2–4–8–10	qd.
D 28	100 muscovite	1–2–4–8–10	qd.
D 29	303 quartz	1–2–3–4	qd.
D 30	100 biotite	1–2–4–6–8	qd.
D 31	100 muscovite	1–2–4–6–8	qd.
D 32	230 quartz	1–2–3–4–5	qd.
D 33	100 biotite	1–2–4–6–8–10	gr.
D 34	100 muscovite	1–2–4–8–12	gr.
D 35	237 non-selected quartz	1–2–3–4–5	gr.
D 36	175 selected quartz	1–2–3–4–5	gr.
D 37	150 biotite	1–2–4–6–8	qd.
D 38	100 muscovite	1–2–4–6–8	qd.
D 39	153 selected quartz	1–2–3–4	qd.
D 40	208 non-selected quartz	1–2–3–4–5	qd.
D 41	108 biotite	1–2–4–6–8	grd.
D 42	100 muscovite	1–2–4–6–8	grd.
D 43	205 non-selected quartz	1–2–3–4–5–6	grd.
D 44	112 selected quartz	1–2–3–4–5	grd.
D 45	100 biotite	1–2–4–8–12	hbn.
D 46	100 biotite	1–2–6–9–12–15	sch.
D 47	100 muscovite	1–2–6–10–14	sch.
D 48	38 muscovite	Porphyroblasts	sch.

* qd.—quartz diorite; hbn.—hornblende gneiss; gr.—granite; gg.—gray gneiss; sch.—schist.

Contours in per cent	Rock* type	Locality
-2–4–8–12–16	gg.	Sec. 11, T. 20 N., R. 19 E. Elmore County
-2–3–4–5	gg.	
-2–4–8–10	qd.	
-2–4–8–10	qd.	Sec. 11, T. 20 N., R. 18 E. Elmore County
-2–3–4	qd.	
-2–4–6–8	qd.	
-2–4–6–8	qd.	Sec. 18, T. 22 N., R. 18 E. Coosa County
-2–3–4–5	qd.	
-2–4–6–8–10	gr.	
-2–4–8–12	gr.	Sec. 17, T. 22 N., R. 20 E. Coosa County
-2–3–4–5	gr.	
-2–3–4–5	gr.	
-2–4–6–8	qd.	
-2–4–6–8	qd.	Sec. 2, T. 21 N., R. 20 E. Coosa County
-2–3–4	qd.	
-2–3–4–5	qd.	
-2–4–6–8	grd.	
-2–4–6–8	grd.	Rockford Coosa County
-2–3–4–5–6	grd.	
-2–3–4–5	grd.	
-2–4–8–12	hbn.	Sec. 6, T. 21 N., R. 19 E. Coosa County
-2–6–9–12–15	sch.	
-2–6–10–14	sch.	Sec. 7, T. 22 S., R. 7 E. Clay County
Porphyroblasts	sch.	

*—hornblende gneiss; gr.—granite; grd.—granodioschist.



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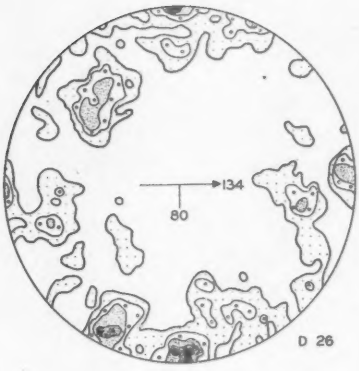
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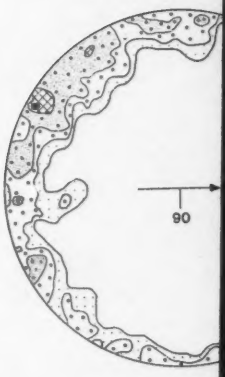
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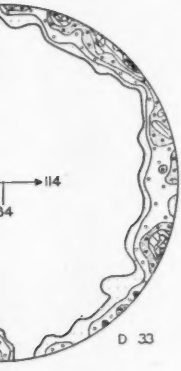
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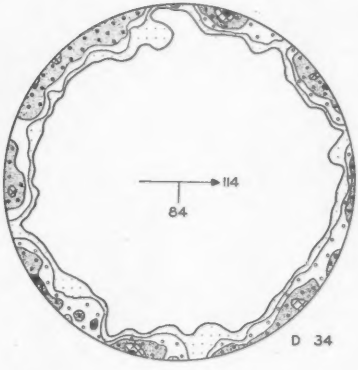
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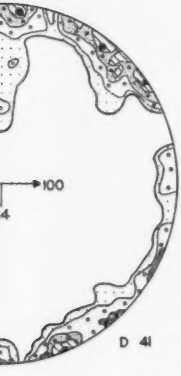
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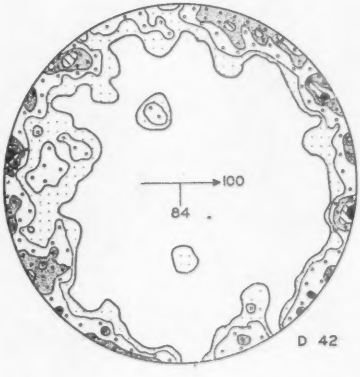
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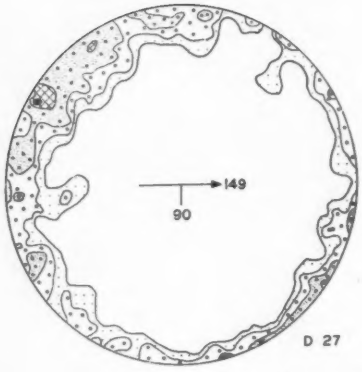

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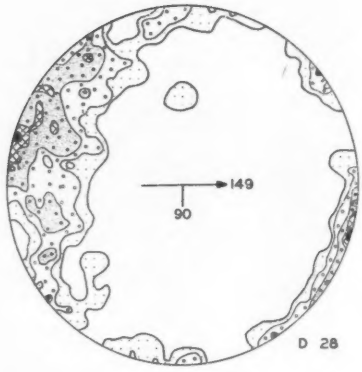

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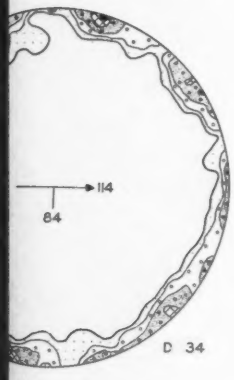

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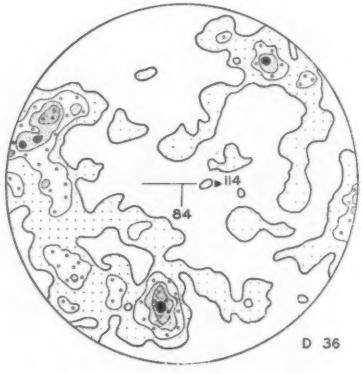

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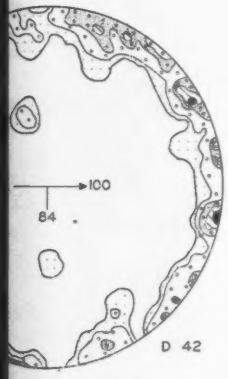

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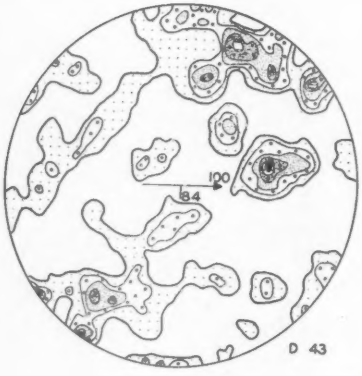

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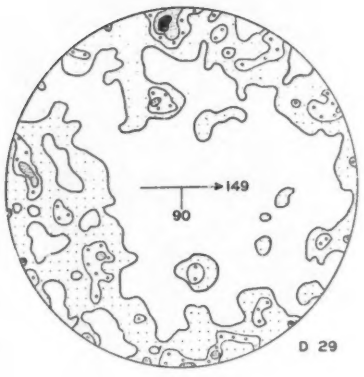

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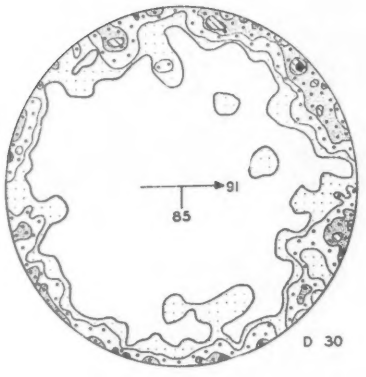

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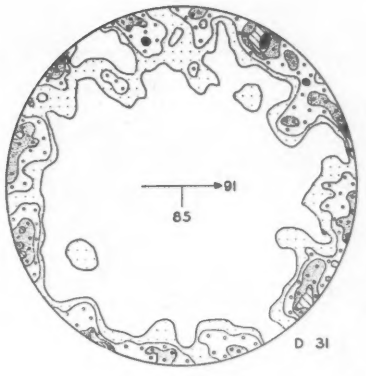

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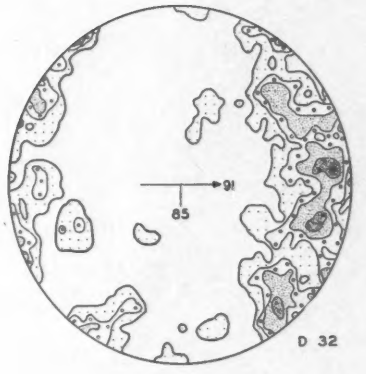

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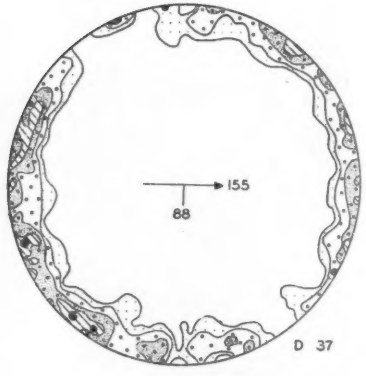

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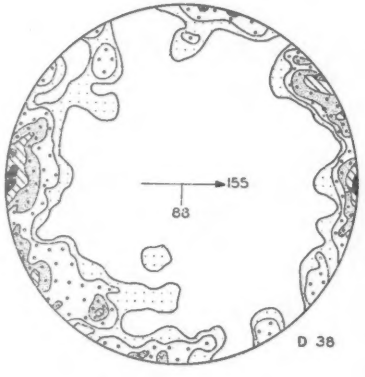

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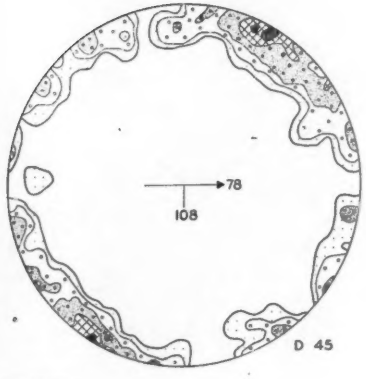

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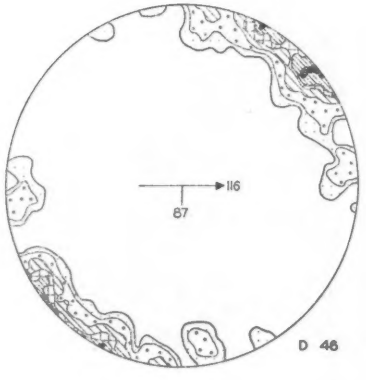

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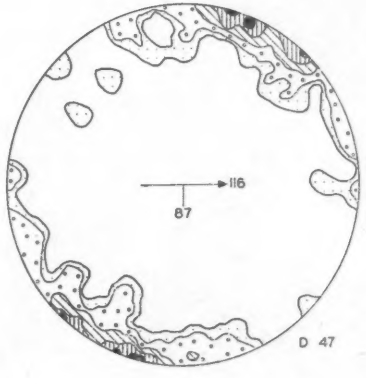

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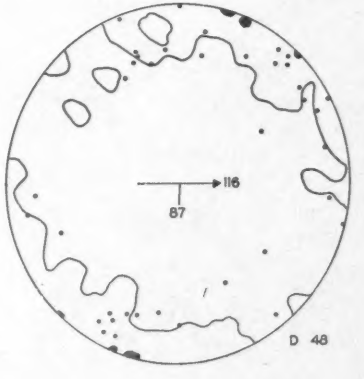

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IAGRAMS

moving plastic mass would be near its contact with the relatively stationary wall. The quartz diorite is more competent than the schist with which it is in contact, and after the quartz diorite became solid the schists yielded more readily to regional deformation than did the quartz diorite. The foliation of the schists parallels the the contacts at many places, and parallelism of flow structures to contacts is characteristic of orientations due to magmatic flow (Balk, 1937). However, the regular trend of all structures in the quartz diorite and their parallelism and similarity to wall-rock structures indicate they are not entirely the result of igneous flow (Stenzel, 1936; Barth, 1936).

The many small disconformable contacts and the northeast strike of the schist foliation at these contacts suggest regional deformation during and after intrusion. The great aggregate length of concordant contacts relative to discordant contacts shows the control of the path of intrusion by earlier wall-rock structures.

Many dikes are discordant, and their structures intense and conformable. Failure of dike walls to influence the dike magma is due to the elongation in *b* (Balk, 1937). Younger dikes in the same outcrop with older dikes contain less well-developed structures.

It therefore seems evident that the Pinckneyville quartz diorite was intruded after much of the folding of the metamorphic rocks but that regional deformation was in effect for some time after emplacement. This deformation is responsible for many of the prominent structures now observed. Small faults, joints, and mylonite zones are structures of the solid phase, as is the fracture cleavage in the schists along the western margin.

The quartz diorite at Hog Mountain is an outlier of the Pinckneyville complex, and Park (1935) believes it was emplaced after much of the regional deformation had taken place but that regional forces were also operative after intrusion. McCaskey (1908) also recognized that the Hog Mountain body was intrusive into the schists and believes that as the rock is not gneissoid it is later than the metamorphism. The principal fractures in which the ore is found at Hog Mountain are roughly normal to the regional strike and in general are not continuous into the schist walls. Wisser (1939) considers that these fractures are tension joints and that they are the same as the cross joints of the main complex. The veins in the fractures show evidence of deformation during mineralization (Wisser).

The Pinckneyville complex is thought to have been intruded as a mush of crystals with some interstitial fluid material, or possibly some parts of the complex came in as hot plastic material. The intrusive material was forced into the flexible schists following, in general, the foliation planes and other prominent northeast-trending planar structures. The power of small dikes to force aside the schists and to deform them locally so that their schistosity parallels the contact (Fig. 27) reflects the intrusion of larger bodies.

The schists were intruded by many bodies of igneous material ranging in size from small dikes (Fig. 21) to large masses such as that at Hog Mountain, following many planes rather than one distinct channel. The material apparently was injected into the schists much as grease can be squeezed up through the fingers. Near the upper boundary of the magma and along the margins there were many small sheets of ig-

neous material. As intrusion proceeded the schists were forced farther and farther apart, and the center of the complex was mostly igneous material. There may also have been some slight displacement of the schists upward.

This concept of the mode of intrusion is suggested by the contact relations along and near part of Socapatoy Creek, by the intercalation of gneiss and schist on the

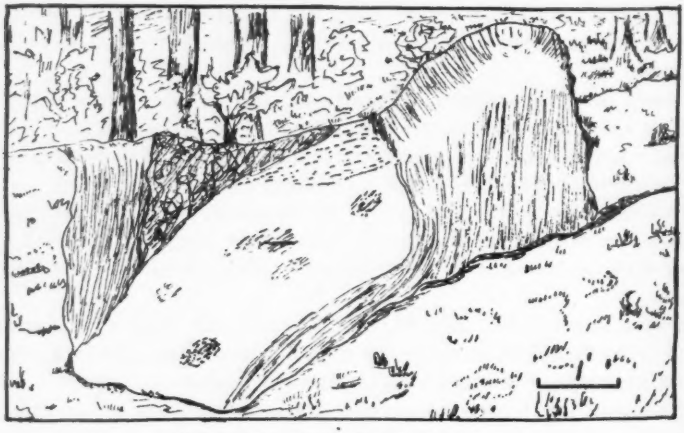

FIGURE 27.—*Granodiorite dike in schist*
West bank of Socapatoy Creek, Coosa County.

western margin, and by the many schist septa and inclusions at the northeast end of the complex. In the area shown in Plate 4, Socapatoy Creek has cut through a northeast-trending ridge underlain by schist, but in the creek where it cuts through the ridge only granodiorite is exposed. Small dikes and apophyses of granodiorite extend up into the schists on the banks for short distances and then terminate. Throughout this small area schist is much more abundant at higher altitudes than at lower altitudes. The contact relations here are interpreted as showing the roof or part of the roof of a body of granodiorite. The random arrangement of many inclusions and the folding of schist septa suggest some turbulence in the flow of the magma which might be expected at the roof. The alternation of granodiorite and schist to the west represents many small sheets of granodiorite which probably become larger at depth until they form one large mass of granodiorite.

The intercalation of gneiss and schist along the western margin of the complex (Fig. 28; Pl. 1) is due to the same structure as that in the Socapatoy Creek area. The schist septa within the complex at the northeast end are probably roof pendants or large inclusions.

The general direction of intrusion of the Pinckneyville quartz diorite may have been from south to north at a moderate angle upward. Park's (1935) description of the Hog Mountain outlier suggests that the major axis of that body pitches gently southwest. Within the complex the number and size of the septa and inclusions decrease from northeast to southwest indicating there is less distance between the

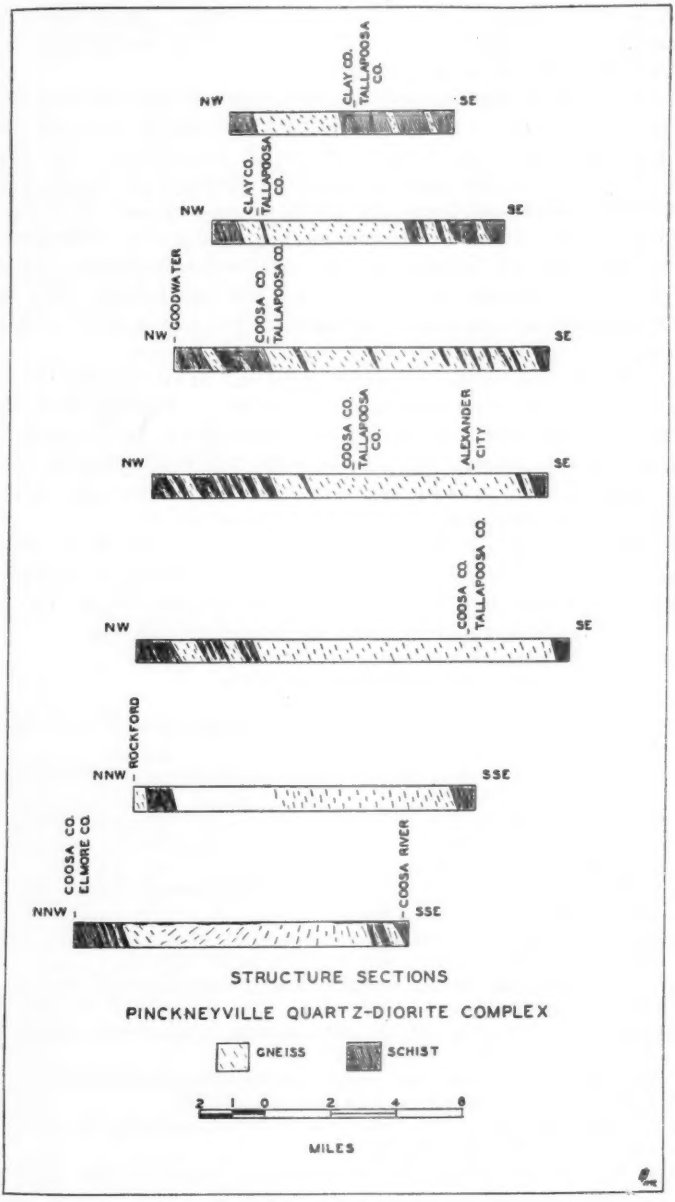

FIGURE 28.—*Structure sections*

Through Pinckneyville quartz diorite complex.

roof and the present surface to the north than to the south. Furthermore, the northeast boundary of the complex is an intrusive contact, while the southwest boundary is unknown because of the Cretaceous overlap.

The relative time of intrusion with respect to wall-rock deformation is fairly certain. The intensity of the structures and their similarity to wall-rock structures leaves little doubt that the Pinckneyville is not posttectonic. The discordant contacts, the evidence of structural control of intrusion, and the structures of inclusions indicate that regional forces were active before intrusion. Intrusion during folding explains all structures. The conformable contacts often mask the intrusive character, but this is to be expected in syntectonic intrusions (Högbom, 1928).

The petrofabric analyses also agree with the field observations. The mica and quartz orientations are due in part to magmatic flow and in part to regional deformation.

The age of the Pinckneyville complex is uncertain. It intrudes the Ashland mica schist and the Wedowee formation (Adams, *et al.*, 1926). The Ashland is considered by some to be Algonkian, and the Wedowee is thought to be Cambrian to Carboniferous. Adams believed the Pinckneyville was post-Carboniferous. Jonas (1932) considered granites in Georgia and the Carolinas along the strike of the Pinckneyville to be late Paleozoic.

The syntectonic origin of the Pinckneyville quartz diorite and the similarity of structures of this area to Appalachian structures and to other late Paleozoic intrusives (Williams and Billings, 1938; Hershey, 1937) suggest that the Pinckneyville quartz diorite was probably intruded near the close of the Paleozoic.

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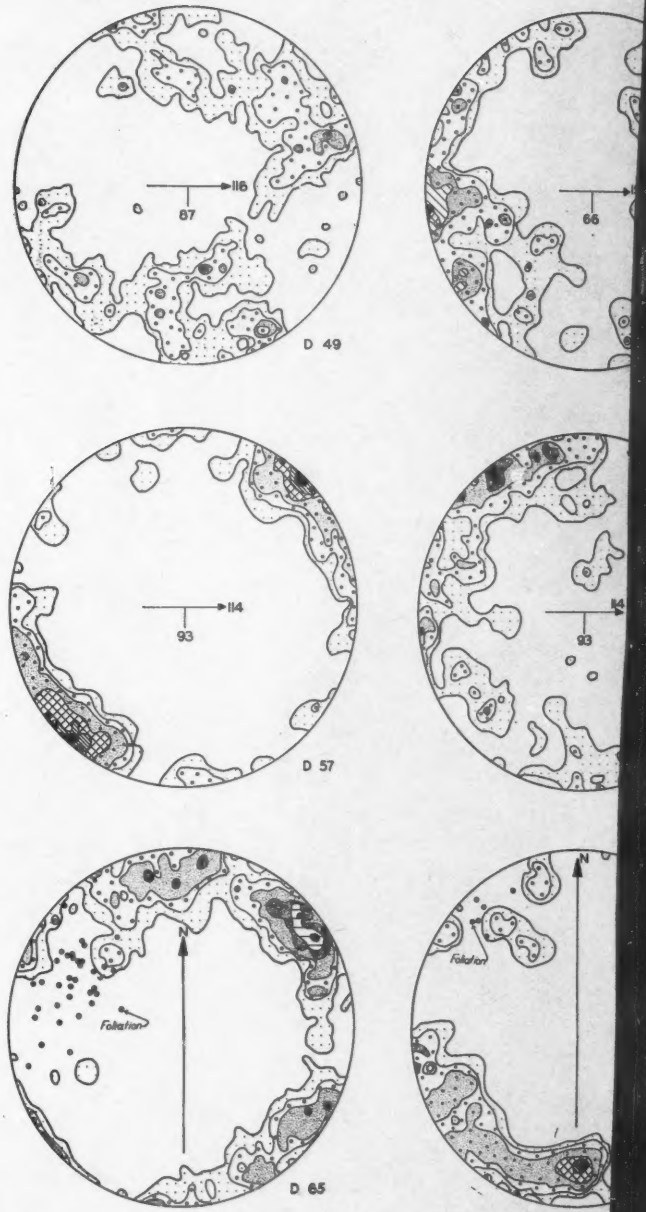
Diagram number	Number and type of grains measured	Contours in per cent	Rock type*	Locality
D 49	250 quartz	1–2–3–3½	sch.	Sec. 7, T. 22 S., R. 7 E. Clay County
D 50	104 biotite	1–2–4–6–8–9	qd.	
D 51	101 muscovite	1–2–4–6–8	qd.	Sec. 4, T. 22 S., R. 7 E. Clay County
D 52	250 quartz	1–2–3–4	qd.	
D 53	103 biotite	1–2–6–10–14–18–22	sch.	
D 54	103 muscovite	1–2–6–10–14–18–22	sch.	Sec. 21, T. 22 N., R. 21 E. Tallapoosa County
D 55	250 quartz	1–2–3–4	sch.	
D 56	100 biotite	1–2–4–8–12–14	grd.	
D 57	100 muscovite	1–2–4–8–12–14	grd.	Sec. 21, T. 22 N., R. 21 E. Tallapoosa County
D 58	200 quartz	1–2–3–4–4½	grd.	
D 59	Composite diagram of major maxima of mica diagrams. D 15 not plotted. 30 biotites, 27 muscovites.			
D 60	Composite diagram of 4 per cent or higher quartz maxima and 3, 3½ per cent maxima from D 49. 82 points. Contours, per cent, 1–2–4–6–8.			
D 61	207 joints; sec. 10, T. 21 N., R. 20 E., Coosa County. Contours, per cent, 1–2–3–4–5. Cherty quartzite, brecciated.			
D 62	100 joints; sec. 35, T. 21 N., R. 19 E., Coosa County. Contours, per cent, 1–2–4–6. Schist.			
D 63	107 joints; sec. 35, T. 21 N., R. 19 E., Coosa County. Contours, per cent, 1–2–4–6–8–10. One fourth mile west of D 62. Gneiss.			
D 64	148 joints; Socapatoy Creek. Contours, per cent, 1–2–4–6–8. Schist.			
D 65	309 joints; Socapatoy Creek. Contours, per cent, 1–2–3–4–5–6. Granodiorite.			
D 66	99 joints; sec. 35, T. 22 N., R. 20 E., Coosa County. Contours, per cent, 1–2–4–8–12–14. Gneiss.			
D 67	224 joints; downstream from contact on Elkahatchee Creek. Contours, per cent, 1–2–4–6–8. Schist.			
D 68	257 joints; 250 feet upstream from contact on Elkahatchee Creek. Contours, per cent, 1–2–4–6–8–9. Gneiss.			
D 69	182 joints; 450 feet upstream from D 68, Elkahatchee Creek. Contours, per cent, 1–2–4–8–12–14. Gneiss.			
D 70	337 joints; upstream from D 69, Elkahatchee Creek. Contours, per cent, 1–2–4–8–10. Gneiss.			
D 71	329 joints; Elmore County. Contours per cent, 1–2–4–6–8–10. Gneiss.			

* qd.—quartz diorite; sch.—schist; grd.—granodiorite.



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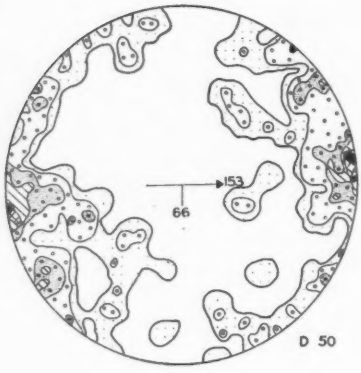

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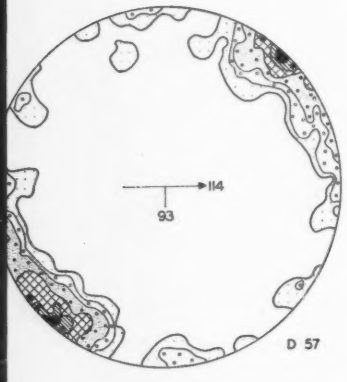

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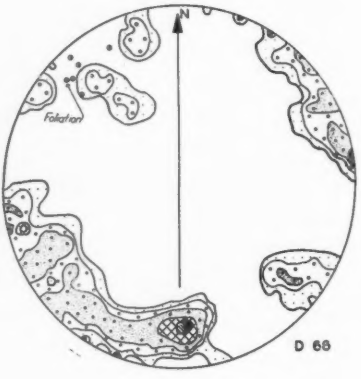

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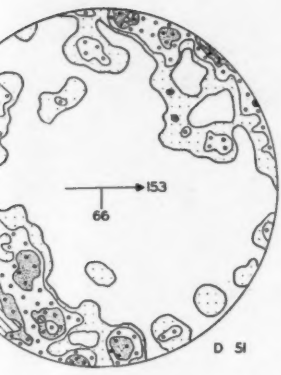

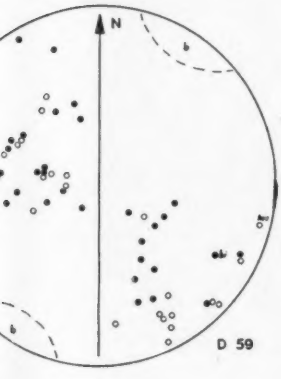

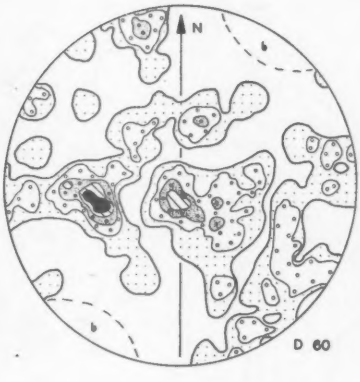

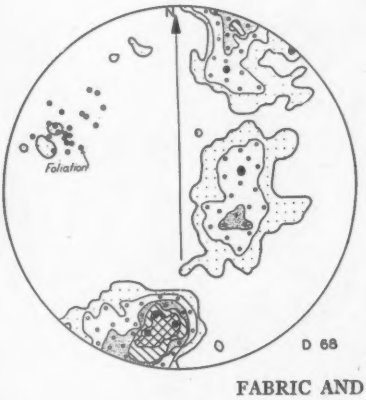

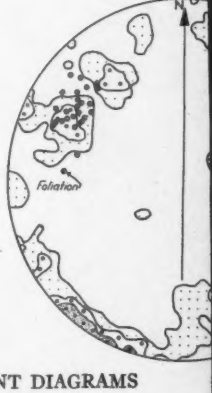

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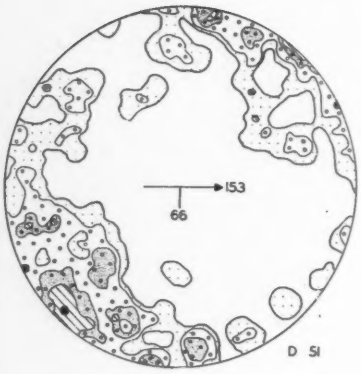

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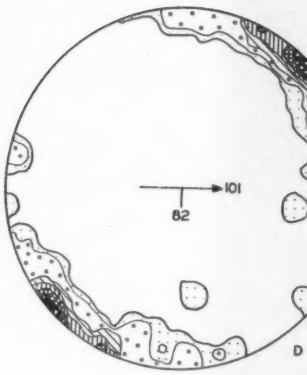

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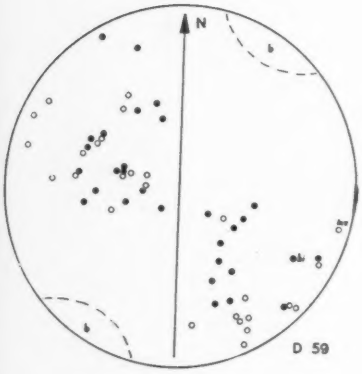

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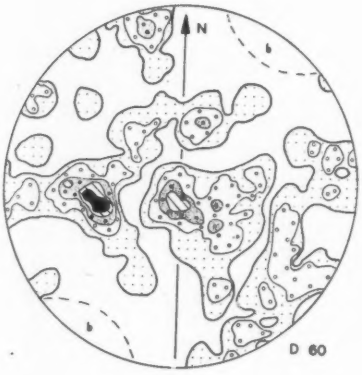

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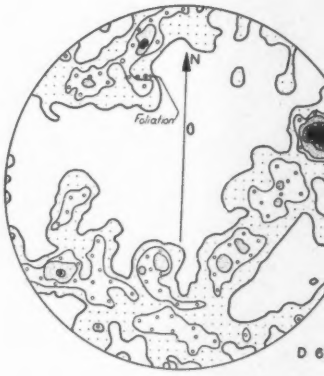

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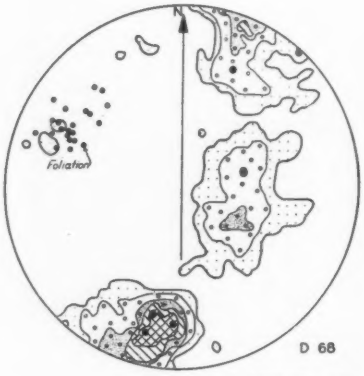

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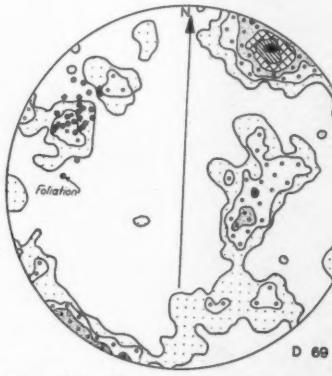

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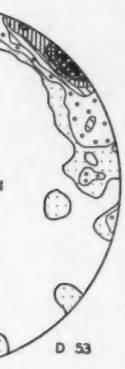

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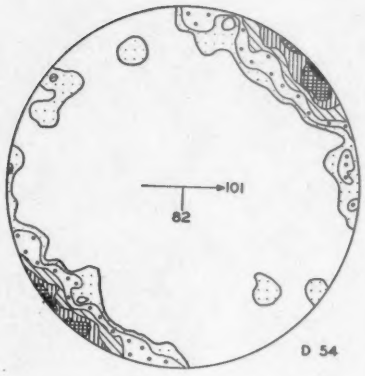

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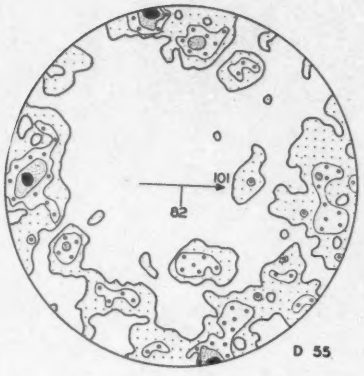

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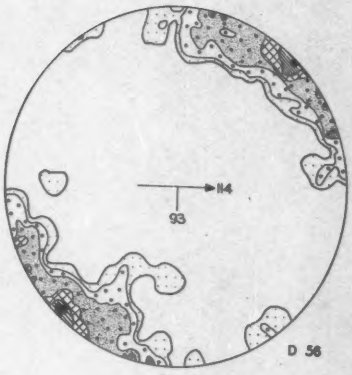

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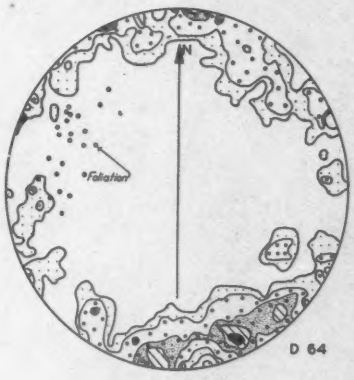

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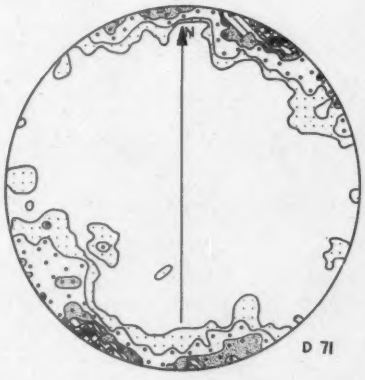
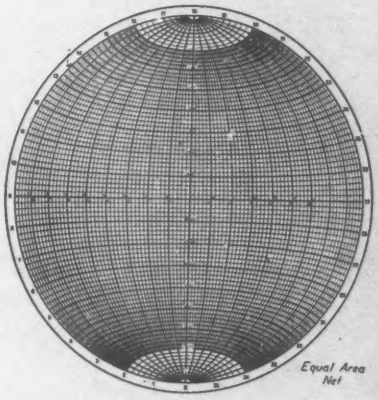
D 70



D 71



Equal Area Net